Maintenance and Sudden Change of a Strong Elevated Ducting Event Associated with High Pressure and Marine Low-Level Jet

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ABSTRACT

Capture of a strong elevated ducting event, especially its maintenance and sudden change, is of great value to airborne radar to achieve its beyond-the-line-of-sight detection. However, the knowledge is not easily accessible over the open ocean and hence very rare. During the Air–Sea Interaction Survey (ASIS) over the western North Pacific (WNP) in May 2016, a strong elevated ducting event with a long-life period and sudden change in its evolution was observed. Measurements from the ASIS, images from the Himawari-8 satellite, reanalysis data from the ECMWF, and Weather Research and Forecasting (WRF) model, were used to analyze the maintenance and sudden change of this strong ducting event, together with the model performance on simulating it. The results showed that the maintenance of strong elevated ducts, with their tops ranging from 750 to 1050 m and average strength of approximately 38 M units, was caused by a strong dry air mass capping over the wet marine atmospheric boundary layer (MABL), together with the subsidence inversion associated with high pressure. The WRF model performs well in simulating them. However, a sudden increase in duct height with a slight decrease of strength was recorded by the subsequent GPS radiosonde, which was finally contributed to the mechanical turbulent inversion and hydrolapse associated with the marine low-level jet (MLLJ). The height of the maximum horizontal wind speed ($U_{\text{nh}}$) of the MLLJ corresponds well with the bottom of the trapping layer. However, these jet-relevant ducts are generally weak and it is difficult to accurately simulate them by using the mesoscale numerical model, since the wind-shear produced eddies are too small to be properly parameterized.

Key words: atmospheric ducts, elevated ducts, marine atmospheric boundary layer (MABL), low-level jet, numerical simulation

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1. Introduction

Propagation of electromagnetic (EM) waves emitted by radar or communication device depends strongly on atmospheric refractive conditions related to the vertical thermodynamic structure of the atmospheric boundary layer (Bean and Dutton, 1968; Turton et al., 1988; Wang et al., 2018). The trapped refraction is a special type of anomalous propagation where the EM wave energy from a transmitting source is trapped within a shallow atmospheric layer known as the ducting layer, or simply called as the atmospheric duct (Alappattu et al., 2016). When trapped within a duct, EM waves propagate to extended ranges to realize “beyond-the-line-of-sight” propagation. A corresponding reduction in the strength of the EM signal occurs above or below the duct, and is referred to as a “radar hole” (Brooks et al., 1999). Atmospheric ducts have a significant influence on the performance of radar, communication device, and weaponry, which enhance their theoretical significance and practical value in understanding the formation of and mechanism of changes in atmospheric ducts for the real-time analysis, numerical simulations, and application to forecasting.

Atmospheric ducts can be defined by the vertical gradient of the modified refractivity ($M$), as below:

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\[ M = \frac{77.6}{T} (p + 4810 e/T) + z/R \times 10^6, \]  

(1)

where \( T \) (K) is the atmospheric temperature, \( p \) (hPa) is the atmospheric pressure, \( e \) is the water vapor pressure (hPa), \( z \) (m) is the height above sea level or ground, and \( R \) is the mean radius of the earth. The trapping layer is the layer with \( dM/dz < 0 \), and duct properties like the duct top, duct strength, and duct thickness, are defined by using the \( M \) profile (Fig. 1). According to the definition of \( M \) in Eq. (1), the sharp vertical decrease of moisture and/or temperature inversion is conductive to the formation of atmospheric ducts (Bean and Dutton, 1968; Turton et al., 1988). Ducts can be classified into three categories: surface, elevated, and evaporation ducts (Kulessa et al., 2017). Due to the particularity of the marine atmospheric environment, the probability of occurrence of a duct is much higher than that on the land. Ocean surface evaporation that leads to the sharp vertical decrease of moisture often results in evaporation ducts, because of which they are permanent or semi-permanent over the sea (Atkinson et al., 2001; Ding et al., 2015; Zhang et al., 2017). In addition, the tropical and subtropical marine atmosphere boundary layers (MABLs) are often capped with an inversion layer (IL), wherein the temperature inversion and sharp vertical decrease of moisture with height facilitate the formation of elevated ducts (Haack and Burk, 2001; Liu and Liang, 2010). Therefore, atmospheric ducts over the sea are the focus of the relevant research because of their high probability of occurrence and long-term existence (Atkinson and Zhu, 2006; Zhang et al., 2016; Kulessa et al., 2017).

Studying atmospheric ducts requires the radiosonde data at a high vertical resolution, but the conventional radiosonde data over the sea have been rarely recorded. To obtain the propagation loss of EM waves as well as concurrent atmospheric refraction conditions over the sea, many series of large-scale joint observational experiments have been carried out, such as the Wallops-2000 Microwave Propagation Measurement Experiment (Wallops 2000; Thompson and Haack, 2011), the Rough Evaporation Duct experiment (RED; Anderson et al., 2004), Sea Breeze 2009 experiment (Garrett et al., 2011), Tropical Air–Sea Propagation Study (TAPS; Kulessa et al., 2017), and Coupled Air–Sea Processes and EM ducting Research (CASPER; Wang et al., 2018). These experiments have yielded plenty of meteorological data and propagation measurements that provide reliable support for research on atmospheric ducts. Moreover, the mesoscale numerical model has played an increasingly important role in the research and prediction of atmospheric ducts in recent years, and can effectively compensate for observational data in the insufficient spatial and temporal resolution (Haack et al., 2010; Liu et al., 2012; Telišman Prtenjak et al., 2015; Zhao et al., 2016; Hu et al., 2018; Ulate et al., 2019).

An extremely strong process of elevated ducts was captured by high-resolution GPS radiosondes in the Air–Sea Interaction Survey (ASIS) over the western North Pacific (WNP) conducted in May 2016. The process was stable and long-last, and has a great research value. This study analyzed the evolution and cause of the extremely strong process of elevated ducts by using the measurements from ASIS, images from the Himawari-8 satellite, and reanalysis data from the ECMWF (0.125° × 0.125°). The process was simulated by using the Weather Research and Forecasting (WRF) model, and the results were evaluated and used to reveal their spatial and temporal evolutionary characteristics.

Fig. 1. The schematic diagram of an elevated duct.
2. Observational data

ASIS over the WNP was conducted in May 2016. The meteorological data were drawn from two sources—GPS radiosondes and surface stations (measured at the bow mast of the research ship, hereinafter referred to as AUTO)—and were processed by quality control software. GPS radiosondes were launched from the research ship. The temperature, humidity, pressure, and winds were measured by AUTO at a temporal resolution of 20 Hz, and the hourly AUTO data at 10-min average intervals was used in this study. The observation data from 0000 UTC 25 May to 0000 UTC 28 May were selected for analysis. In this period, the ship traveled from south to north along 160°E across 11 latitudes (29°–40°N) and a total of 13 GPS radiosondes were launched.

3. Analysis of cause and evolution of ducts

3.1 Observations

Among the 13 GPS radiosondes launched from 0000 UTC 25 May to 0000 UTC 28 May, seven consecutive observations of strong elevated ducts were recorded (from 0100 UTC 26 May to 1100 UTC 27 May; Fig. 2d). As the observational location moved northward over time, the duct strength from GPS radiosondes data gradually weakened, decreasing from 43.2 to 27.3 M units. The height of the duct top showed a non-monotonic change of first decreasing and then increasing. It is interesting to note that as the observational location moved northward, the height of the duct top suddenly increased by 700 m to reach 1650 m from 0500 to 1100 UTC 27 May (Fig. 2d). Profiles of the temperature and humidity (Figs. 2a, b, e) showed a typical structure of the convective MABL, i.e., a prominent temperature inversion and a sharp vertical decrease in humidity above the top of the mixing layer where the strong elevated ducts appeared. The strong elevated ducts vanished with the disappearance of temperature inversion and sharp vertical decrease in humidity. This strong elevated duct process has an important research value as one of the few observational cases in open oceans, together with its long-life period and dramatic evolution, which was comprehensively captured owing to the movement of the research ship.

3.2 Cause analysis

In this section, ECMWF reanalysis (ERA)-Interim (0.125° × 0.125°) data, observational data from ASIS, and images from the Himawari-8 satellite are used to analyze the evolution and cause of the strong elevated ducts, especially the cause of sudden rise in height of the duct top and the accompanying weakening in its strength at 1100 UTC 27 May.

3.2.1 No strong elevated ducts observed

From 0000 UTC 25 May to 1800 UTC 25 May, no strong elevated ducts (> 10 M units) were observed. As shown in Fig. 3a, there was a high-pressure anticyclonic system with its center at about 150°E behind a trough (near 170°E), according to the 850-hPa geopotential height at 0000 UTC 25 May. The northwest airflow behind this trough guided the dry air mass to the south, combined with the subsidence inversion and dehumidification due to the high-pressure system, so a strong dry air mass was accumulated in the area almost coincided with the high-pressure system. This dry air mass at 850 hPa covered on the wet marine boundary layer, leading to a sharp vertical decrease in humidity, which was favorable to the formation of atmospheric ducts. During this period (Fig. 3a), however, the observational locations (marked by the cyan star in Fig. 3) were just outside the southeast boundary of the dry air mass, and were not controlled by the high pressure, because of which there was no inversion and sharp vertical decrease in humidity in the boundary layer. No ducts were observed either. With the eastward movement of the entire synoptic situation and the northward movement of the observational location, the observational location shifted to the southern boundary of the high-pressure anticyclonic system until 1800 UTC 25 May (Fig. 3b). But before that, it was still not under the control of the high pressure and dry air mass. Thus, no strong elevated ducts were observed during this period.

3.2.2 Continuously strong elevated ducts observed

From 0100 UTC 26 May to 0500 UTC 27 May, strong elevated ducts were observed by six GPS radiosondes in succession as the observational location moved northward (Fig. 2d). The height of the duct top first dropped and then increased, and the variation range was 750–1050 m. The duct strength slightly decreased, and the average strength was about 38 M units. The large-scale synoptic situations at 850 hPa were shown in Figs. 3c–e during this period, and the observational locations were controlled by the high pressure and strong dry air mass. Due to the subsidence inversion induced by the high pressure and coverage of strong dry air mass over the wet MABL, temperature inversion and sharp vertical decrease in humidity were observed (Figs. 2a, b), and strong elevated ducts were recorded in succession (Fig. 2d).

The reasons why the height of the duct top exhibited a
non-monotonic trend of decreasing and then increasing were further analyzed. As shown in Figs. 3c–e, the observational locations moved gradually from the southern edge to the southwestern section and then to the western edge of the strong dry air mass, with their own northward movement and the synoptic situation’s eastward

Fig. 2. Profiles of the (a) potential temperature ($\theta$; K), (b) water vapor mixing ratio ($q_v$; g kg$^{-1}$), (c) wind speed (WS; m s$^{-1}$), (d) modified refractivity ($M$; M units), and (e) temperature ($T$; °C) measured by GPS radiosondes. Colors represent the dates on which the data were recorded.
Fig. 3. Horizontal distributions of the specific humidity (colored; g kg$^{-1}$), geopotential height (black contour; gpm), wind (vector; m s$^{-1}$), and temperature (blue contour; °C) at 850 hPa from the ERA-Interim dataset (0.125° × 0.125°) at (a) 0000 UTC 25 May, (b) 1800 UTC 25 May, (c) 0000 UTC 26 May, (d) 1200 UTC 26 May, (e) 0600 UTC 27 May, (f) 1200 UTC 27 May, and (g) 1800 UTC 27 May 2016. The observational location is marked by the cyan star.
movement. As to the distance from the high-pressure center, the observational locations that gradually moved northward were first close to and then far from the center. This trend could be confirmed by the change in sea level pressure measured by AUTO (Fig. 4c). The sea level pressure generally showed a trend of increasing first and then decreasing in this period, corresponding to the observational location’s first approaching and then gradually moving away from the high-pressure center. The peak value of sea level pressure appeared at 1200 UTC 26 May, which meant that the observational location was the closest to the center of high pressure at this time. In addition, the wind speed profile below 3 km measured by the GPS radiosonde (Fig. 2c) showed that the wind speed at 1200 UTC 26 May was the lowest overall in this period, which was consistent with the fact that the high-pressure center was nearly static. Wan and Sun (2010) have shown that strong inversion inhibits the development of the boundary layer. Because subsidence near the center of high pressure is generally the strongest, inversion is also the strongest, and thus the height of the boundary layer was at its lowest (Figs. 2a, b), as was the height of the duct top (Fig. 2d) at 1200 UTC 26 May. This also explained why the height of the duct top de-

![Graph](https://example.com/graph.png)

**Fig. 4.** (a) Temperature ($T$; °C), (b) water vapor mixing ratio ($q_v$; g kg$^{-1}$), (c) sea-level pressure (SLP; hPa), (d) 10-m wind speed (WS; m s$^{-1}$), and (e) wind direction (WD; °) from the AUTO (black) and the WRF simulation (red).
increased first and then increased. In addition, the duct strength observed in this period slowly weakened, which was also related with the movement of observational locations relative to the high-pressure system (Figs. 3d, e). The southerly airflow prevailed on the west side of the high pressure, resulting in the suppression of subsidence due to the advection of warm and moist air mass. In addition, as the entire synoptic situation moved eastward, the observational location was nearly beyond control by the strong dry air mass, which also weakened the vertical gradient in terms of humidity. The duct strength thus slowly weakened.

3.2.3 Sudden increase in height of the duct top and disappearance of strong elevated ducts

The observational location gradually deviated from the control by the high-pressure system and strong dry air mass, and was affected by the warm and moist southwesterly airflow from 1200 UTC 27 May (Figs. 3f, g). Two favorable conditions for duct formation mentioned in the above analysis, temperature inversion and sharp vertical decrease in humidity, are not satisfied, which explains why the disappearance of strong elevated ducts at the observational location began at 1200 UTC 27 May (Fig. 2d). However, a strong elevated duct with the height (duct top) of 1650 m and strength of 27.3 M units was observed at 1100 UTC 27 May, although the large-scale situation was not conducive to its formation.

Profiles of the temperature and humidity at 1100 UTC (2130 LT) 27 May showed a weak stable layer at the lower level (200 m below) of the MABL (Figs. 2a, b, e; note that the lower layer was unsaturated and $\gamma < \gamma_d$), and there was a nocturnal low-level jet (NLLJ) near the top of the stable layer at 270 m (Fig. 2c). The maximum wind speed was 15 m s$^{-1}$, and its presence is a typical feature of a nocturnal stable boundary layer. However, another marine low-level jet (MLLJ) was also observed in the horizontal wind speed profile at the height of 1600 m (Fig. 2c), and corresponded precisely to the height where the strong elevated duct appeared. The MABL is commonly a convective boundary layer (CBL), in which a thermal turbulent attenuation layer usually exists between the main body of CBL (i.e., the mixing layer) and upper free atmosphere (see the schematic diagram in Fig. 5a). The original temperature profile ($\text{AFB}$; the red line) and dew-point temperature profile ($\text{ABC}$; the blue line) can convert to the $\text{AFB'}B$ and $\text{ABC'}$ after turbulent mixing, and then the temperature inversion ($\text{F'B}$) and sharp decrease of moisture versus height (i.e., hydrolapse and $\text{B'C}$) are formed just within the turbulent attenuation layer. This is a well-known explanation for the formation of turbulent inversion, and the detailed mechanism can be attributed to the dry adiabatic law of the unsaturated air parcel. Theoretically speaking (as in Fig. 5b), a sheared-produced (mechanical) turbulent attenuation layer can also form just above the height of the maximum horizontal wind speed ($U_{\text{in}}$) of the MLLJ, yielding a temperature inversion and hydrolapse in the similar way. On this occasion, the height of $U_{\text{in}}$ should correspond to the bottom of the trapping layer. Taking the case at 1100 UTC 27 May for instance, the maximum wind shear of this MLLJ reached 8 m s$^{-1}$ (100 m)$^{-1}$, and $U_{\text{in}}$ of the MLLJ was located at about 1600 m (Fig. 2c), accompanied by a strong temperature inversion (almost 6°C; Fig. 2e) and hydrolapse (Fig. 2b) just above this altitude, and the trapping layer was truly bottomed at about 1600 m as expected (Fig. 2d). Another example was at 0000 UTC 25 May, the height of the MLLJ $U_{\text{in}}$ (at about 2500 m) also corresponded to the bottom of a weak trapping layer (Fig. 2d), and the mechanical turbulent inversion had a thickness of about 100 m just above this level. Similar examples included 1800 UTC 27 May and 0000 UTC 28 May at about 2650 and 2700 m, respectively.

Therefore, the sudden increase in the height of the duct top and decrease in its strength at 1100 UTC 27 May were related to the change of the formation cause. The strong elevated ducts continuously observed before were caused by the subsidence inversion and strong dry air mass control in the high-pressure system. However, the elevated duct observed at 1100 UTC 27 May was due to the mechanical turbulent attenuation associated with the MLLJ. The height of $U_{\text{in}}$ corresponded well with the bottom of the trapping layer. These jet-relevant ducts are generally weak. It seems that these ducts can be strengthened when the wind shear of the LLJ is strong [here the maximum is about 8 m s$^{-1}$ (100 m)$^{-1}$] and the mechanical turbulent attenuation is combined with a thermal attenuation. However, it must be noted that the mechanical turbulent attenuation associated with the MLLJ only provides a formation probability. Whether the jet-relevant ducts are really appeared or not is also determined by the synoptic situations. For instance, the capped subsidence inversion and strong dry air advection tend to enhance them, whereas the moisture increase with height (sometimes occurred in a foggy or rainy day) is hostile to their formation, such as the strong MLLJ at the height of 200 m at 1800 UTC 27 May.

The cause of formation of the MLLJ was then analyzed by using the GPS radiosondes (Figs. 2a–c), synoptic situations from ERA-Interim data (Figs. 3f, g), and visible cloud images from the Himawari-8 satellite (Figs.
First of all, from 1100 UTC (2130 LT) to 1800 UTC (0430 LT) 27 May, the low-level boundary layer was a weak stable layer with NLLJ just above its top at nearly 270 m, which is a typical structure for the nocturnal boundary layer above the land. NLLJ usually forms after sunset, reaches its strongest in early morning hours, and decreases during the day. The formation of NLLJ is usually associated with the inertial oscillations of the wind field that is inherent in a nocturnal stable boundary layer (Li and Shu, 2008). However, such a low-level stable boundary layer was not common over the open ocean, and it strengthened unexpectedly at 0000 UTC (1030 LT) 28 May. Profiles of the temperature and humidity (Figs. 2a, b, e) showed a strong inversion and vertical increase of humidity at lower level below 120 m, which generally corresponded to the occurrence of sea fog or precipitation. Combined with the analysis of the visible cloud map (Figs. 6a, b), it was clear that a cold front accompanied with an extratropical cyclone during this period passed at an eastward moving speed of about 0.83 longitude per hour (92 km h$^{-1}$). Therefore, the passage of the cold front was the main reason for formation of the stable boundary layer and NLLJ over the ocean. In addition, from the large-scale situation (Fig. 3g), abundant water vapor was transported by the warm and moist southwestly airflow and precipitation did exist at the observational location at 0000 UTC 28 May, evidenced by the 3-h cumulative precipitation from the ERA-Interim dataset (Figs. 6c, d). It also can be inferred that the surface frontal had already passed through the observational location at 0000 UTC 28 May, and therefore precipitation occurred in the area under the control of the cold air mass. This is why there was an NLLJ and temperature inversion just above the sea surface, but no surface ducts were observed. The MLLJ occurring between 1600 and 2600 m during this period may also be associated with the strong baroclinicity near the frontal zone and hence an intense thermal wind. Figures 3f and 3g showed that the isotherm was indeed dense near the observational location.
4. Numerical simulations

4.1 Model description and configuration

WRF version 3.6.1 was used for the simulation. The WRF model is nonhydrostatic, three dimensional, and fully compressible, with the higher-order closure and a terrain-following coordinate scheme. It is state-of-the-art mesoscale forecast model designed for both research and operational applications, and has been thoroughly tested and successfully used in many studies.

The major physical settings included the Yonsei University (YSU) boundary layer scheme, Kain–Fritsch cumulus parameterization scheme, Rapid Radiative Transfer Model for Global Climate Models (RRTMG) shortwave–longwave radiation scheme, and Lin microphysics scheme. The period of the simulation was 0000 UTC 24 May 2016 to 0000 UTC 28 May 2016, with the first 24 h treated as model spin-up. Three nested domains (D01, D02, and D03) were introduced to the model with horizontal resolutions of 45, 15, and 5 km, respectively. D01 covered roughly 0–50°N, 130°E–170°W. From the surface to 50 hPa, the model had 64 unevenly spaced vertical layers, and thickness of each was 50 m below 1.6 km in the boundary layer. The initial and boundary conditions of the simulations were obtained from the 1° × 1° NCEP Final (NCEP-FNL) operational global analysis data.

4.2 Results

4.2.1 Model results evaluated against observations

The simulation results were evaluated against measurements from AUTO and GPS radiosondes. The gridded model data were interpolated to the time and location of each observational datum. For comparisons with AUTO, data for the bottom layer (about 8 m) in D03 were used. The temperature, water vapor mixing ratio, and wind direction from the simulation agreed well with the observed values (Figs. 4a, b, c); the trend of simulated sea level pressure was also in good agreement with the observation, but was higher by about 2.5 hPa overall (Fig. 4c); the 10-min average wind speed measured by
AUTO was volatile but the model successfully simulated the two peaks of wind speed at 1300 UTC 25 May and 2200 UTC 27 May (Fig. 4d).

Figure 7 shows that the duct strength and duct top height diagnosed from 13 GPS radiosondes and the corresponding simulation data (if there were multilayered ducts, the strongest was used for comparison). The simulation successfully reproduced the strong elevated ducts from 0100 UTC 26 May to 0500 UTC 27 May, which was caused by the subsidence inversion associated with high pressure and the sharp vertical decrease in humidity related to strong dry air mass. The simulated height (duct top) and strength of the duct agreed well with the observations. However, the model failed to predict the jet-relevant elevated ducts, or predicted them with large deviations. The simulated duct at 1100 UTC 27 May was lower and weaker (Fig. 7). Because the scale of the wind-shear produced eddies are usually very small, the vertical resolution of the GPS radiosonde (about 5 m) is sufficient to recognize them, whereas the vertical resolution of the mesoscale numerical model is currently not fine enough.

### 4.2.2 Simulated characteristics of evolution of strong elevated ducts

Since the WRF model performs well in simulating the strong elevated ducts associated with high pressure and strong dry air mass control, we further reveal their comprehensive spatial and temporal distribution based on the simulation results. As shown in Fig. 8, the main type of this strong ducting event was the elevated duct that was long-last throughout the simulation period. The strong elevated ducts (> 10 M units) moved eastward following the high pressure with a decrease in height along the meridional direction to the north and closer to the high-pressure center along the zonal direction (Figs. 8c, f, i), which was consistent with the observations mentioned before. The extremely strong elevated ducts (> 30 M units) mainly appeared at the southern half of the high-pressure system, which remains an interesting issue for our subsequent study.

### 5. Conclusions and discussion

A strong elevated ducting event over the open ocean, with both a long-life period and a sudden change in its evolution, was recorded during ASIS over the WNP in May 2016. The measurements from ASIS, images from the Himawari-8 satellite, and data from ERA-interim were used to analyze the maintenance and sudden change of this strong ducting event. Numerical simulations, as well as the model performance validation was also conducted based on this event by using the WRF model,
since the single-point observation could hardly reflect the overall ducts conditions in a given region.

The results showed that the maintenance of the strong elevated ducts, with their tops ranging from 750 to 1050 m and average strength of approximately 38 M units, was caused by a strong dry air mass capping over the wet MABL, together with the subsidence inversion associated with high pressure. The WRF model performed well in simulating them and these strong ducts showed persistent existence throughout the simulation period. They moved eastward following the high pressure with a decrease in height both to the north along the meridional direction and closer to the high-pressure center along the zonal direction. Interesting to note that the extremely strong elevated ducts (> 30 M units) mainly appeared at the southern half of the high-pressure system, which remains a potential issue for our future study.

However, a sudden increase in duct height with a slight decrease of strength was recorded by the subsequent GPS radiosonde, which was finally contributed to the mechanical turbulent inversion and hydrolapse associated with the MLLJ. The formation mechanism has been depicted schematically in Fig. 5. A sheared-produced (mechanical) turbulent attenuation forms just above the height of $U_{mh}$ of the MLLJ, yielding a temperature inversion and hydrolapse across this turbulent attenuation layer (Fig. 5b). It is similar to the well-known explanation for turbulent inversion (Fig. 5a), according to the dry adiabatic law of the unsaturated air parcel. As a result, the height of $U_{mh}$ corresponds well to the bottom of the trapping layer for the jet-relevant ducts. It is important to note that the mechanical turbulent attenuation associated with the MLLJ only provides a formation probability. Whether these jet-relevant ducts are really appeared or not also determined by the synoptic situations. Generally, these ducts are weak, and it is difficult to accurately simulate them by using the WRF model, since the scale of the wind-shear produced eddies is too small to be properly parameterized, and hence the large eddy simulation (LES) may be a good solution in the future.

This paper is only a preliminary study to shed a light to the phenomenon of jet-relevant ducts. The threshold of the wind shear for a strong jet-relevant duct, combined effect of mechanical and thermal turbulent attenuations, and other reinforcement effects such as the capped subsidence inversion and strong dry air advection, are still necessary to be investigated by a broader dataset and

Fig. 8. Simulated horizontal distributions of the (a, d, g) duct type; (b, c, h) duct strength (colored; M units), geopotential height (black contour; gpm), and wind (vector; m s$^{-1}$) at 850 hPa; (c, f, i) duct top height (colored; m), sea surface pressure (black contour; hPa), and 10-m wind (vector; m s$^{-1}$) at 0100 UTC 26 May, 2300 UTC 26 May, and 1100 UTC 27 May 2016, respectively. The black star represents the observational location.
hence worthy of a future study.

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REFERENCES

Alappattu, D. P., Q. Wang, and J. Kalogiros, 2016: Anomalous propagation conditions over eastern Pacific Ocean derived from MAGIC data. Radio Sci., 51, 1142–1156, doi: 10.1002/2016RS005994.

Anderson, K., B. Brooks, P. Caffrey, et al., 2004: The RED experiment: An assessment of boundary layer effects in a trade winds regime on microwave and infrared propagation over the sea. Bull. Amer. Meteor. Soc., 85, 1355–1366, doi: 10.1175/BAMS-85-9-1355.

Atkinson, B. W., and M. Zhu, 2006: Coastal effects on radar propagation in atmospheric ducting conditions. Meteor. Appl., 13, 53–62, doi: 10.1017/S1350482705001970.

Atkinson, B. W., J. G. Li, and R. S. Plant, 2001: Numerical modeling of the propagation environment in the atmospheric boundary layer over the Persian Gulf. J. Appl. Meteor., 40, 586–603, doi: 10.1175/1520-0450(2001)040<0586:NMOTOP>2.0.CO;2.

Bean, B. R., and E. J. Dutton, 1968: Radio Meteorology. Dover Publications, New York, 435 pp.

Brooks, I. M., A. K. Goroch, and D. P. Rogers, 1999: Observations of strong surface radar ducts over the Persian Gulf. J. Appl. Meteor., 38, 1293–1310, doi: 10.1175/1520-0450(1999)038<1293:OOSROM>2.0.CO;2.

Ding, J. L., J. F. Fei, X. G. Huang, et al., 2015: Development and sensitivity experiments. J. Meteor. Res., 29, 467–481, doi: 10.1007/s13351-015-3238-4.

Garrett, S. A., D. E. Cook, and R. E. Marshall, 2011: The Seabreeze 2009 experiment: Investigating the impact of ocean and atmospheric processes on radar performance in the Bay of Plenty, New Zealand. Wea. Climate, 31, 82–99.

Haack, T., and S. D. Burk, 2001: Summertime marine refractivity conditions along coastal California. J. Appl. Meteor., 40, 673–687, doi: 10.1175/1520-0450(2001)040<0673:SMRCAC>2.0.CO;2.

Haack, T., C. G. Wang, S. Garrett, et al., 2010: Mesoscale modeling of boundary layer refractivity and atmospheric ducting. J. Appl. Meteor. Climatol., 49, 2437–2457, doi: 10.1175/2010JAMC2415.1.

Hu, H., J. F. Fei, J. L. Ding, et al., 2018: Mechanism analysis and numerical simulation of strong marine typhoon duct caused by super typhoon “Lupit” (2009). Acta Meteor. Sinica, 76, 620–634, doi: 10.11676/qxb.2018.024. (in Chinese)

Kulessa, A. S., A. Barrios, J. Claverie, et al., 2017: The Tropical Air–Sea Propagation Study (TAPS). Bull. Amer. Meteor. Soc., 98, 517–537, doi: 10.1175/BAMS-D-14-00284.1.

Li, J., and W. J. Shu, 2008: Observation and analysis of nocturnal low-level jet characteristics over Beijing in summer. Chinese J. Geophys., 51, 360–368, doi: 10.3321/j.issn.0001-5733.2008.02.008. (in Chinese)

Liu, G. Y., S. H. Gao, Y. M. Wang, et al., 2012: Numerical simulation of atmospheric duct in typhoon subsidence area. J. Appl. Meteor. Sci., 23, 77–88, doi: 10.3969/j.issn.1001-7313.2012.01.009. (in Chinese)

Liu, S. Y., and X. Z. Liang, 2010: Observed diurnal cycle climatology of planetary boundary layer height. J. Climate, 23, 5790–5809, doi: 10.1175/2010JCLI3552.1.

Telisliman Prenjak, M., I. Horvat, I. Tomažić, et al., 2015: Impact of mesoscale meteorological processes on anomalous radar propagation conditions over the northern Adriatic area. J. Geophys. Res. Atmos., 120, 8759–8782, doi: 10.1002/2014JD022626.

Thompson, W. T., and T. Haack, 2011: An investigation of sea surface temperature influence on microwave refractivity. The Wallops-2000 experiment. J. Appl. Meteor. Climatol., 50, 2319–2337, doi: 10.1175/JAMC-D-10-05002.1.

Turton, J. D., D. A. Bennetts, and S. F. G. Farmer, 1988: An introduction to radio ducting. Meteor. Mag., 117, 245–254.

Ulate, M., Q. Wang, T. Haack, et al., 2019: Mean offshore refractive conditions during the CASPER East field campaign. J. Appl. Meteor. Climatol., 58, 853–874, doi: 10.1175/JAMC-D-18-0029.1.

Wan, J., and J. N. Sun, 2010: Impact of capping inversion on development of convective boundary layer: Evaluating the parameterization of entrainment rate through large-eddy simulation. Scientia Meteor. Sinica, 30, 715–723, doi: 10.3969/j.issn.1009-0827.2010.05.020. (in Chinese)

Wang, Q., D. P. Alappattu, S. Billingsley, et al., 2018: CASPER: Coupled Air–Sea Processes and Electromagnetic Ducting Research. Bull. Amer. Meteor. Soc., 99, 1449–1471, doi: 10.1175/bams-d-16-0046.1.

Zhang, Q., K. D. Yang, and Y. Shi, 2016: Spatial and temporal variability of the evaporation duct in the Gulf of Aden. Tellus A, 68, 29792, doi: 10.3402/tellusa.v68.29792.

Zhang, Q., K. D. Yang, and Q. L. Yang, 2017: Statistical analysis of the quantified relationship between evaporation duct and oceanic evaporation for unstable conditions. J. Atmos. Oceanic Technol., 34, 2489–2497, doi: 10.1175/JTECH-D-17-0156.1.

Zhao, Q. Y., T. Haack, J. McLay, et al., 2016: Ensemble prediction of atmospheric refractivity conditions for EM propagation. J. Appl. Meteor. Climatol., 55, 2113–2130, doi: 10.1175/JAMC-D-16-0033.1.