



國立臺灣大學理學院大氣科學系

碩士論文

Department of Atmospheric Sciences

College of Science

National Taiwan University

Master's Thesis

TAHOPE SOP4期間之準線性對流系統：

都卜勒雷達分析下的對流特徵與演變

A Quasi-Linear Convective System during the TAHOPE

SOP 4: Convection Characteristics and Evolution from

Doppler Radar Data Analysis

劉仔茗

Yu-Ming Liu

指導教授：楊明仁 博士

Advisor: Ming-Jen Yang Ph.D.

中華民國 114 年 7 月

July 2025

國立臺灣大學碩士學位論文 口試委員會審定書

本論文係劉仔茗君（學號R12229001）在國立臺灣大學大氣科學學系、所完成之碩士學位論文，於民國 114 年 6 月 30 日承下列考試委員審查通過及口試及格，特此證明

口試委員：

楊明仁 (簽名)

(指導教授)

郭鴻基

宋偉國

劉清煌

廖宇慶

楊明仁

系主任、所長

游政谷 (簽名)

Acknowledgements

碩士兩年的生活過得很快，一一回顧的話或許可以寫300頁心得跟教學了，我也不是感性的人，所以這邊就不多寫些什麼，簡短來吧。

感謝楊老師這兩年的指導，獲益良多。謝謝茂正學長、九令學長、曜竹學長、英璋學長、尚恩學長、炯恩學長、品瑞、瑞均、彥葦、晴葳，這些實驗室小夥伴們，在我研究遇到大大瓶頸的時候，跟我一起討論科學、解決問題，在休息時一起歡樂聊比賽、聊天南地北，一起出遊逛貓空，還有好多好多的研究生活回憶！真的很感謝實驗室小夥伴們！以後要多約約吃晚餐OuOb。謝謝愛我的家人們，給我最大的支持跟鼓勵，聽我分享研究生活，給我人生建議。最後，期許自己能夠永遠樂觀往下走，辛苦啦！



中文摘要



本研究探討 2022 年 6 月 29 日於 TAHOPE (Taiwan-Area Heavy Rain Observation and Prediction Experiment) 第四次特別觀測期間(SOP4)所觀測到的一個準線性對流系統 (QLCS) 之演變與特徵。該系統在弱綜觀系統與垂直風切較弱的環境中發展，儘管如此，系統仍展現出短暫的弓形狀回波結構與地面風速增強的現象，這些特徵通常較常出現在強風切環境中。透過 NCAR S-Pol 都卜勒雙偏極化雷達資料、地面氣象站觀測以及 SAMURAI 雷達風場反演技術，本研究分析了該對流系統的結構演變、渦度生成與微物理過程。

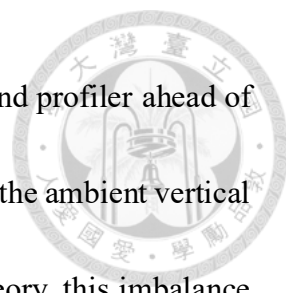
該對流系統逐漸發展為一條長度約 90 公里的 QLCS，並形成短暫的弓形狀回波，其伴隨較強的降水與冷池現象。此弓形回波的生成可歸因於冷池所引發的水平浮力梯度，此梯度產生水平渦度，進而被低層上升氣流傾斜並拉伸為垂直渦度偶極結構。隨著系統分裂，對流系統向北移動並通過 S-Pol 雷達站上空，於新屋氣象站觀測到地面風速增強與風向轉變。由對流胞所伴隨的後到前 (Rear to Front; RTF) 氣流與下沉氣流，可能為導致此地面風速增強與風向變化的主因。根據地面氣壓差與對流系統前方由新屋風剖儀所提供的環境垂直風切資料計算顯示，此事件中的冷池強度大於環境垂直風切。依據 Rotunno-Klemp-Weisman (RKW) 理論，這種不平衡可以解釋系統隨後的減弱與消散現象。對流系統的移動主要受到新對流胞發展位置的控制，而新對流胞的生成除了受到海風輻合作用的影響外，亦與具有較高相當位溫 (θ_e) (具有較大潛在不穩定度) 的熱力條件有關。雙偏極化雷達分析 (包括隨高度變化的頻率分布圖 CFAD) 顯示降水微物理特性隨對流強度及系統演變階段而變化，並觀測到 RTF 氣流通過低層時亦會影響其微物理過程。

English Abstract



This study investigates the evolution and characteristics of a quasi-linear convective system (QLCS) observed on 29 June 2022 during the TAHOPE (Taiwan-Area Heavy Rain Observation and Prediction Experiment) Special Observation Period 4. The system developed under weak synoptic forcing and low vertical wind shear. Despite this, it exhibited transient bow-shaped echo structures and enhanced surface winds, which are typically associated with stronger shear environments. Using dual-polarization Doppler radar data from the NCAR S-Pol radar, surface station observations, and wind retrieval techniques (SAMURAI), this study analyzed the convective system structural evolution, vorticity development, and microphysical processes.

This convective system evolved into a 90-km long QLCS featuring a short-lived bow-shaped echo with enhanced precipitation and cold pool. The formation of the bow-shaped echo was attributed to cold-pool-induced horizontal buoyancy gradients, which generated horizontal vorticity that was subsequently tilted and stretched by low-level updrafts to form vertical vorticity couplets. As the system split, the convective system moved northward, passing over the S-Pol radar site and producing intensified surface winds and directional shifts at the Xinwu station. The rear-to-front (RTF) flow with downdrafts from the convection, likely contributed to the observed surface wind strengthened and direction changed. Based on pressure differences derived from surface



stations and ambient vertical wind shear derived from the Xinwu wind profiler ahead of the convective system, it showed that the cold pool was stronger than the ambient vertical wind shear. According to the Rotunno–Klemp–Weisman (RKW) theory, this imbalance can explain the subsequent weakening and dissipation of the convection. The movement of the convection was governed by the location of new convective cell development, which was influenced not only by sea-breeze convergence but also by thermodynamically favorable regions with higher θ_e (greater instability). Dual-polarization radar analysis, including Contoured Frequency by Altitude Diagrams (CFADs), revealed changes in precipitation microphysics associated with size sorting, evaporation, and stratiform bright band features. These microphysics varied from convective intensity and the stages of convection evolution, and were also associated with observed rear-to-front (RTF) flow passing through low-level layers, influencing microphysical processes along its path.

Table of Contents



Oral Defense Approval Form.....	i
Acknowledgments.....	ii
Abstract (in Chinese).....	iii
Abstract (in English).....	iv
Tabel of Contents.....	vi
List of Tables.....	viii
List of Figures.....	ix
Chapter 1 Introduction.....	1
Chapter 2 Data and Method	7
2.1 Surface Observations.....	7
2.2 Radar Observations and Wind Field Retrieval.....	9
Chapter 3 Case Overview.....	11
3.1 Synoptic condition.....	11
3.2 Evolution of Convective System.....	12
3.3 Surface Observations.....	13
Chapter 4 Bow-shaped echo.....	15
4.1 Formation and Evolution.....	15
4.2 Mechanisms of Bow-shaped Echo.....	17

Chapter 5 Movement of convection and strong surface wind.....	22
5.1 Formation and Evolution.....	22
5.2 Rear-to-Front Flow and Bounded Weak Echo Region.....	23
5.3 Surface Wind Intensification.....	25
5.4 System Dissipation.....	26
Chapter 6 Dual-Polarization Parameter Analysis.....	31
6.1 Z_H - Z_{DR} Difference Space.....	31
6.2 RHI CFADs during Bow-shaped Echo.....	33
6.3 RHI CFADs in Stratiform Region.....	35
6.4 RHI CFADs of Rear Inflow.....	36
Chapter 7 Summary.....	38
Appendix.....	42
Reference.....	48
Tables.....	54
Figures.....	56



List of Tables

Table 1. CAPE and vertical wind shear from Hsinchu during TAHOPE.....	54
Table 2. TAHOPE 2022 IOP and SOP (adapted from Yang et al., 2023).....	54

List of Figures

- Figure 1.** (a) Distribution of surface observation instruments and radar sites in the study area during TAHOPE. (b) The configuration of observational instruments used in the field experiment is illustrated in Fig. 1, adapted from 楊 等人 (2023). 56
- Figure 2.** (a) 500-hPa geopotential height, wind vectors, and wind speed (shaded) at 08 LST on 29 June 2022, derived from the ERA5 reanalysis. (b) The corresponding 850-hPa geopotential height, wind vectors, and equivalent potential temperature (θ_e). (c) 850-hPa synoptic analysis chart. (d) 200-hPa synoptic analysis chart. (e) HIMAWARI IR color image for Asia area. (f) 00 UTC sounding launched by the CSU in Hsinchu. The shaded area represents the CAPE. The wind rose in the top-right corner displays vertical wind shear vectors: red and purple arrows indicate 10 m–2 km and 10 m–5 km shear, respectively. 57
- Figure 3.** Surface wind observations from ground-based stations at (a) 0730 LST, (b) 0830 LST, and (c) 0930 LST. 58
- Figure 4.** Low-level water vapor content observed by the MicroPulse Differential Absorption Lidar (MPD) and converted to vapor mixing ratio (units: g/kg) using the atmospheric density derived from Hsinchu radiosonde observations. 58
- Figure 5.** Composite radar reflectivity from the S-Pol at 2-km altitude from 1448 to 1912 LST. The blue star indicates the location of the S-Pol radar site. Panels: (a) 1448, (b) 1536, (c) 1548, (d) 1625, (e) 1712, (f) 1812, and (g) 1912 LST. 59
- Figure 6.** Major surface observation stations along the convective system's south-to-north propagation path, including Taichung, Qingshui, Miaoli, Hsinchu, and Xinwu. The panel at the bottom right shows the geographic locations of these stations. The time series plots on the left display, from top to bottom, are temperature, pressure (with the semi-diurnal tide removed), wind direction, and wind speed respectively. 60

Figure 7. Composite reflectivity in units of dBZ at 2 km altitude and horizontal storm-relative wind (SRWIND) from 1536 to 1712 LST. The blue star indicates the location of the S-Pol radar. Pink contours mark regions where vertical velocity exceeds 4 m s^{-1} at 2-km height, highlighting areas of strong low-level updrafts. Panel: (a) 1536, (b) 1548, (c) 1600, (d) 1625, (e) 1637, (f) 1648, (g) 1700, (h) 1712 LST. 61

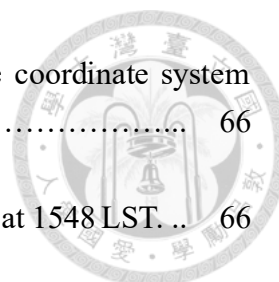
Figure 8. Temperature anomaly (in units of K) overlaid with 45 dBZ radar reflectivity contours and 10-minute accumulated rainfall from 1536 to 1712 LST. The size of the pink circles is proportional to the magnitude of accumulated rainfall; the largest circle represents a maximum accumulation of 20 mm per 10 minutes. Panel: (a) 1536, (b) 1548, (c) 1600, (d) 1625, (e) 1637, (f) 1648, (g) 1700, (h) 1712 LST. 62

Figure 9. Storm-relative wind speed (units: m s^{-1}) and wind vectors overlaid with 45 dBZ radar reflectivity contours. Pink contours indicate regions where wind speed increases. The blue star marks the location of the S-Pol radar. Pink contours also denote areas where vertical velocity exceeds 4 m s^{-1} at 2-km height. Panel: (a) 1536, (b) 1548, (c) 1600, (d) 1625. 63

Figure 10. (a) Domain-averaged vorticity budget terms, with different colored lines representing individual contributions from each term. (b)-(e) vorticity (shaded; units: 10^{-3} s^{-1}) and the sum of the stretching and tilting terms from the vertical vorticity tendency equation (contours; units: 10^{-5} s^{-2}). Solid contours indicate positive tendencies (0.5, 1.0, and 2.0), while dashed contours represent negative tendencies (-0.5 , -1.0 , and -2.0). Panels (b)–(e) correspond to 1536, 1548, 1600 and 1625 LST, respectively. 64

Figure 11. (a) Schematic illustration of the coordinate system rotation. (b) Conceptual diagram showing the tilting and stretching of horizontal vorticity. ... 65

Figure 12. Cross-sectional analysis of the convective system at 1536 LST. (a) Location of cross section AA' overlaid on the composite radar reflectivity, indicating the system's position at this time. (b) Wind field along the AA' cross section, with the vertical component exaggerated by a factor of two. Contours represent radar reflectivity, with shading applied for values exceeding 45 dBZ. (c) Shaded colors represent radar reflectivity; contours denote regions where $K_{DP} \geq 1^\circ \text{ km}^{-1}$. (d) Horizontal vorticity projected



	onto the rotated cross-sectional plane (following the coordinate system shown in Fig. 11a); contours indicate vertical velocity.	66
Figure 13.	As in Fig. 12, but showing the cross-sectional analysis at 1548 LST. ...	66
Figure 14.	Cross-section of wind direction and speed derived from the Xinwu wind profiler between 1400 and 2000 LST. Wind barbs indicate wind direction and speed, and the background color shading represents wind speed magnitude as shown in colorbar. The red box marks the time period when changes in wind direction and speed begin to appear.	67
Figure 15.	Locations of RHI cross-sections at azimuth angles of 183° and 0°	67
Figure 16.	Same as Fig. 7, but for a different time period. Star indicates the location of S-Pol radar site. Panel: (a) 1637, (b) 1712, (c) 1736, (d) 1800, (e) 1824, (f) 1848, (g) 1912, LST.	68
Figure 17.	Same as Fig. 8, but for a different time period. Panel: (a) 1712, (b) 1724, (c) 1736, (d) 1748, (e) 1800 LST.	69
Figure 18.	RHI radar analysis. (a) PPI image showing the location of the convective system and the orientation of the RHI cross section at 1637 LST. (b) RHI cross section of reflectivity with overlaid contours of $K_{DP} \geq 1^\circ \text{ km}^{-1}$. (c) RHI cross section of radial velocity and wind vectors along the plane of the cross-section, with vertical velocity exaggerated by a factor of 2.	70
Figure 19.	Same as Fig. 18, but at 1648 LST.	70
Figure 20.	Same as Fig. 18, but at 1700 LST.	71
Figure 21.	(a)-(c) Same as Fig. 18, but at 1712 LST. (d) RHI cross-section of ZDR. The dark indigo box highlights the region where the K_{DP} column extends above the melting layer (4–5 km) and where both DBZ and Z_{DR} signals are weak.	71
Figure 22.	Left panels show the 183° RHI cross sections, and right panels show the 0° RHI cross sections. From top to bottom, the panels display radar	

	reflectivity overlaid with areas where $K_{DP} \geq 1^\circ \text{ km}^{-1}$, and radial velocity overlaid with horizontal wind vectors. The time is at 1736 LST.	72
Figure 23.	Same as Fig. 22, but at 1748 LST.	72
Figure 24.	Same as Fig. 22, but at 1800 LST.	73
Figure 25.	Vertical shear of horizontal wind in the lower to mid-troposphere, as derived from the Xinwu wind profiler. The blue line indicates vertical shear between 10 m and 2 km, the orange line represents shear up to 3 km, and the green line shows shear up to 5 km (note that some data are missing).	73
Figure 26.	Temperature anomaly (in units of K) overlaid with 45 dBZ radar reflectivity contours and surface wind from (a)16:37, (b) 16:48, (c) 17:00 to (d)17:12 LST.	74
Figure 27.	Surface equivalent potential (in units of K) temperature from (a)16:37, (b) 16:48, (c) 17:00 to (d)17:12 LST.	75
Figure 28.	(a) Sounding profile from Hsinchu at 00 UTC, (b) sounding profile from Hsinchu at 06 UTC, (c) sounding profile from Makung at 00 UTC, and (d) 925-hPa wind barbs and wind speed (shading).	76
Figure 29.	(a) Definition of the concentric horizontal analysis domain with radii from 30 to 100 km, centered on the S-Pol radar (blue star). The domain is based on the radar's observable altitude coverage from the gridded PPI reflectivity data. (b) Diagram adapted from Kumjian (2014; 2022) to illustrate warm-rain processes, showing only the second quadrant. Different line styles represent different time periods.	77
Figure 30.	Locations of the convective system at 1536 and 1548 LST, along with RHI scans from various azimuthal angles. Concentric semicircles are drawn at 10-km radial intervals.	78
Figure 31.	Median CFADs of radar reflectivity (dBZ) within the 60–90 km range for RHIs between 190° and 205° . The (a)-(d) correspond to 1536 LST, and (e)-(h) correspond to 1548 LST.	79

Figure 32.	Same as Fig. 31, but for K_{DP} . (a)-(d) correspond to 1536 LST, and (e)-(h) correspond to 1548 LST.	80
Figure 33.	Same as Fig. 31, but for ZDR. (a)-(d) correspond to 1536 LST, and (e)-(h) correspond to 1548 LST.	81
Figure 34.	Azimuthal averages of RHIs during the stratiform precipitation period at 18:12 LST: (a)-(c) for DBZ, (d)-(e) for ZDR, and (f)-(i) for RHOHV.	82
Figure 35.	Same distance interval, illustrating distribution differences between the convective and stratiform regions. The RHI azimuths correspond to those shown in Fig. 30b and 30c. Different colored lines represent individual azimuthal directions, while the black star-marked dashed line indicates the mean profile. Panel (a) represents the convective region at 1548 LST, and panel (b) represents the stratiform region at 1812 LST.	83
Figure 36.	Differences observed in the 30–40 km range before and after the arrival of the RTF flow. Different colored lines represent individual azimuthal directions, and the black star-marked dashed line indicates the mean profile. The vertical panels show radial velocity, Z_{DR} , and Z_H in panels (a), (d), and (g), respectively. The horizontal panels represent different times: 1648, 1700, and 1712 LST in panels (a), (b), and (c), respectively.	84
Figure 37.	Conceptual diagram illustrating the formation of a bow-shaped echo.	85
Figure 38.	Schematic summary of key processes during the northward propagation and dissipation of the QLCS.	85

Chapter 1. Introduction

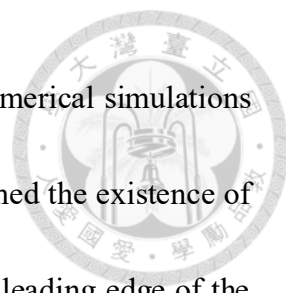


Mesoscale Convective Systems (MCSs) are frequently associated with high-impact weather phenomena such as short-duration heavy rainfall, lightning, and damaging surface winds. These nature hazards pose significant threats to human activities and infrastructure. When MCSs develop into linear structures and exhibit characteristic features such as line echo wave patterns (LEWPs), bow echoes, book-end vortices (BEVs), rear-inflow jets (RIJs), or mesovortex signatures, they are generally classified as Quasi-Linear Convective Systems (QLCSs) (Weisman and Davis 1998; Agee and Jones 2009). Among these weather features, bow echoes, which are characterized by a bulging leading edge and a recessed trailing edge, are commonly associated with intense convection and severe surface wind events (Fujita 1978; Weisman 2001; Agee and Jones 2009).

Previous studies have extensively examined the formation mechanisms and environmental conditions favorable for bow echoes. Both observational and modeling studies indicate that bow echoes typically develop in environments with moderate to strong vertical wind shear in the lower to middle troposphere, along with high convective available potential energy (CAPE) (Rotunno et al. 1988; Weisman 1993, 2001; Weisman and Trapp 2003; Weisman and Rotunno 2004; Wakimoto et al. 2015; Pan and Yang 2022). Ambient moisture can also influence bow echoes development by modulating cold pool

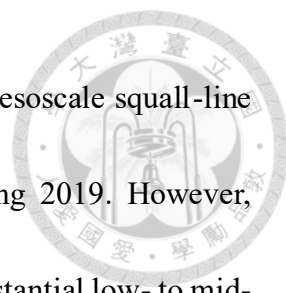
strength, which in turn affects the interaction between the cold pool and the environmental wind shear (James et al. 2006; Mathias et al. 2017). For derecho, studies have shown that under weak synoptic forcing, derecho in environment with larger convective available potential energy (CAPE) values becomes more critical to maintaining system longevity and preventing rapid dissipation (Evans and Doswell III 2001; Coniglio et al. 2004).

Klimowski et al. (2004) showed that bow echoes can originate from various convective modes, including squall lines, supercells, and isolated convection. Their formation is often associated with the merger of convective systems and individual cells. A primary mechanism traditionally recognized for the formation of bow echoes within linear convective systems is the development of a rear-inflow jet (RIJ). The RIJ, as defined by Smull and Houze (1987), is a current of environmental air that penetrates the storm from the rear through the stratiform precipitation region. Both observational and modeling studies have identified the RIJ as a key contributor to the bowing structure of squall lines or organized convective lines (Fujita, 1978; Rotunno et al., 1988; Weisman, 1992, 1993; Grim et al., 2009). Numerical simulations by Weisman (1992, 1993) suggested that the RIJ is driven by enhanced vertical buoyancy gradients generated by a surface cold pool, which induces midlevel pressure perturbations and promotes front-to-rear acceleration of air. Smull and Houze (1987) proposed that a mesoscale pressure perturbation within the stratiform region, in conjunction with another within the



convective region, contributes to the rear-to-front flow structure. Numerical simulations by Yang and Houze (1995) under mixed-phase microphysics confirmed the existence of this rear-to-front (RTF) flow, which extends the rear inflow into the leading edge of the convective system. This interaction leads to the development of a double-core structure, with distinct features in both the trailing stratiform region and the leading convective zone [see Fig. 5b of Yang and Houze (1995)]. Additionally, when ambient horizontal vorticity is tilted and stretched by the convective updraft along the system's leading edge, midlevel book-end or line-end vortices can form (Weisman, 1993; Grim et al., 2009). These vortices contribute to vertical vorticity coupling and may further enhance the RIJ, resulting in a more pronounced bowing of the echo. Aside from environmental vorticity, cold-pool-generated horizontal vorticity within the convective system itself can also contribute to this coupling mechanism (Weisman and Davis, 1998; Wakimoto et al., 2015).

During the warm season in Taiwan, some linear convective systems exhibiting bow-shaped reflectivity structures develop ahead of or along the Mei-yu front. These systems occasionally generate mesoscale vortices during the mature stage of the bow echo, although their overall intensity is generally weaker than those observed in the United States, and they have not been associated with tornado formation (Chen et al., 2012). A few cases have been documented during the winter season; however, they typically occur under strong vertical wind shear exceeding 20 m s^{-1} and are accompanied by significant



baroclinicity (Che et al., 2007). Pan and Yang (2022) examined a mesoscale squall-line system that produced a bow echo as it traversed Taiwan in spring 2019. However, proximity soundings from Magong still indicated the presence of substantial low- to mid-level shear. In general, these case studies suggest that most bow echo events in Taiwan have occurred under relatively strong synoptic forcings, with vertical wind shear commonly exceeding 10 m s^{-1} . The associated convective systems tend to be larger in the horizontal scale and persist over longer durations, in contrast to the QLCS observed on 29 June 2022 during the TAHOPE. This QLCS particular case developed under a weak synoptic background typical of the warm season, with a relatively weak vertical wind shear but high CAPE. Despite the marginal shear environment, the QLCS brought substantial rainfall to the western plains of Taiwan. Specifically, the convective system temporarily developed a bow-shaped reflectivity structure and showed signs of vertical-vorticity coupling. Following the convective system split, the northern portion of the system continued to propagate northward and, during its passage over the S-Pol radar site, produced increases in surface wind speed and directional shifts (Fig. 6). In Taiwan, bowing structures that form under weak shear environments and their subsequent surface impacts remain understudied. In particular, the mechanisms behind bow-shaped echo formation in the absence of a stratiform region differ from those traditionally associated

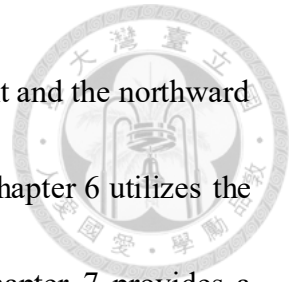
with the presence of a rear-inflow jet (RIJ) as described by Weisman (1992, 1993). These observations raise the following scientific questions:



1. How can a bow-shaped echo briefly develop under weak environmental vertical wind shear ($<10 \text{ m s}^{-1}$), and what physical mechanisms are responsible for its formation?
2. After the convective system split and partially moved offshore, why did the northern segment intensify surface winds during its passage over the S-Pol site and continue to propagate northward?
3. What physical processes governed the system's evolution from its mature stage through dissipation?

To address the aforementioned scientific questions, this study employs the S-Pol Doppler radar data and surface observations combined with wind field retrieval diagnostics to analyze the structural evolution of the convective system at various stages. The sources and evolution of vorticity during the development of the bow-shaped echo are quantitatively examined, supplemented by the Contoured Frequency by Altitude Diagram (CFAD) analyses to investigate changes in precipitation particles and vertical motions. This approach aims to elucidate the coupled dynamical and microphysical processes involved in the event. The thesis is organized as follows: Chapter 2 describes the data and methodology; Chapter 3 presents the case background and environmental conditions; Chapter 4 analyzes the convective bowing development and vertical vorticity

structure; Chapter 5 explores the causes of surface wind enhancement and the northward propagation of the system, as well as the dissipation mechanisms; Chapter 6 utilizes the CFADs to analyze the physical characteristic changes; finally, Chapter 7 provides a comprehensive conclusions.

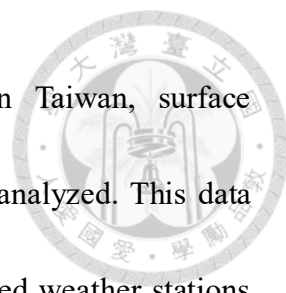


Chapter 2. Data and Methodology



The observational data in this study were primarily obtained from the Taiwan-Area Heavy Rain Observation and Prediction Experiment (TAHOPE, https://exp.pccu.edu.tw/TAHOPE_2022/), conducted from 25 May to 11 August 2022 (楊等人, 2023). During this period, the National Center for Atmospheric Research (NCAR) deployed the S-band polarimetric (S-Pol) radar over the coastal region of Hsinchu. The observation period encompassed multiple convective weather events in the central and northern Taiwan. Detailed observation events can be found in Table 2. The S-Pol radar performed standard Plan Position Indicator (PPI) scans as well as Range-Height Indicator (RHI) scans at various azimuths, providing comprehensive vertical structural information. This radar deployment offered unique insights into the convective characteristics and storm evolution, addressing previous observational limitations over the Taichung, Miaoli, and Hsinchu plains, where the distant Nantun and Shulin C-band radars often had observation coverage gaps due to the range and terrain blockages. In addition to radar data, this study integrates various observational datasets for comprehensive analysis. The location of instruments and radars utilized is illustrated in Fig. 1a. For the complete list and placement of observation instruments, please refer to the instrument layout shown in Fig. 1b, adapted from 楊明仁 (2023).

2.1 Surface Observations



Considering the study area spans the central to northern Taiwan, surface observational data between Changhua and Taoyuan counties were analyzed. This data includes 10-minute interval records from both manual and automated weather stations operated by the Central Weather Bureau (CWB). Furthermore, additional radiosonde data were obtained by Colorado State University (CSU) at the Hsinchu City Government site. These radiosondes were released every six hours at 00, 06, 12, and 18 UTC. Prior to this study, we had used radiosonde data to compute various thermodynamic and dynamic parameters to discuss the environmental characteristics associated with the occurrence or non-occurrence of convective initiation (CI). For details, please refer to appendix.

Notably, this field experiment marked the first application of the MicroPulse Differential Absorption Lidar (MPD) system in Taiwan deployed by NCAR to observe the atmospheric water vapor. The MPD employs a differential absorption lidar (DIAL) technique using two separate laser wavelengths: an absorbing wavelength (online) and a non-absorbing wavelength (offline). By analyzing the ratio of returned signals, the system estimates water vapor (Spuler et al., 2021). The MPD provides high temporal and vertical resolution data, recording measurements every 10 minutes with a vertical resolution of 40 meters, covering altitudes from 200 to 4000 meters above ground level. For this study, quality-controlled MPD data collected in Hsinchu during the TAHOPE fourth Special Observation Period (SOP4) were utilized. These MPD data were combined with

atmospheric density profiles derived from radiosonde measurements to convert water vapor into mixing ratios for further analysis.

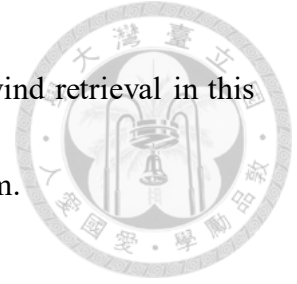


2.2 Radar Observations and Wind Field Retrieval

To analyze the evolution of the QLCS in this case study, high-resolution data from the NCAR S-Pol radar (S-band) and the Nantun and Shulin radars (C-band) from CWB were employed. The convection exhibited a quasi-linear structure with a horizontal extent of approximately 90 kilometers, characteristic of mesoscale β -scale systems (20–200 km; Orlanski (1975)). Key physical properties analyzed in this study include radar reflectivity, radial velocity, and dual-polarization parameters obtained from the S-Pol radar, which were used to examine the evolution of the convection system and its correlation with surface observations.

For the dynamical analysis of wind fields and physical mechanisms within the convective system, this study uses the Spline Analysis at Mesoscale Utilizing Radar and Aircraft Instrumentation (SAMURAI) wind retrieval technique developed by Bell et al. (2012). This method integrates three-dimensional variational data assimilation (3DVAR) with finite element methods to estimate dynamic and thermodynamic fields by minimizing a cost function, yielding the most probable atmospheric state. Cha and Bell (2023) further enhanced this method by incorporating the Immersed Boundary Method (IBM), allowing terrain forcing to be included in the retrieval algorithm, thereby imposing

realistic terrain boundary conditions on the simulated fields. The wind retrieval in this study have a horizontal grid size of 1 km and a vertical size of 0.5 km.



Chapter 3. Case Overview



3.1 Synoptic condition

This study focuses on the SOP4 from the Taiwan-Area Heavy Rain Observation and Prediction Experiment (TAHOPE), which occurred over the central Taiwan on 29 June 2022. By this time, Taiwan had entered the active phase of summertime afternoon convection, when the Mei-yu season was over. At 500 hPa (Fig. 2a), the subtropical high ridge had shifted eastward, resulting in a weak easterly flow. In addition to the mid-level features, the 850 hPa and 200 hPa synoptic charts (Fig. 2c and 2d) showed a tropical disturbance located to the southwest of Taiwan and an upper-level cold-core low to the northeast. According to Wu et al. (1976), stronger upward motion was typically found around the cold-core low. A climatological study by Chen et al. (1988) further noted that, in the region west of 140°E and between 25°–35°N, the southeastern quadrant of cold-core lows tended to exhibit the most unstable conditions and greatest cloud coverage, while the northwestern quadrant had the least amount of cloud cover. This pattern was consistent with satellite image (Fig. 2e), which showed enhanced cloudiness southeast of the cold-core low at 00 UTC. Convective clusters also appeared to the southwest of the cold-core low, located northeast of Taiwan. The tropical disturbance might have contributed to moisture transport toward Taiwan, as cloud clusters were also observed to the island's southwest. Although both the cold-core low and the tropical disturbance

might have enhanced instability and moisture over Taiwan, their influence remained uncertain due to their relatively distant locations and warranted further investigation. However, considering their distance, the synoptic-scale forcing on this day was likely not strong.

The 850 hPa level exhibited relatively moist conditions (Fig. 2b), while radiosonde data (Fig. 2f) indicated a relative dry layer near 500 hPa. The CAPE on this day was approximately 3233.8 J kg^{-1} , indicating a highly unstable atmosphere. However, the vertical wind shear was relatively weak, with wind-speed difference below 10 m s^{-1} in the lower to middle troposphere (Fig. 2f). Surface observations (Fig. 3) showed a transition from calm or land breeze conditions to sea breeze after 0830 LST, indicating the establishment of a sea-land breeze circulation. Data from the MPD (Fig. 4) also revealed a slight increase in mixing ratio by $1\text{--}2 \text{ g kg}^{-1}$ around 0900 LST, reaching a value of approximately 24 g kg^{-1} . This suggests that the inland penetration of the sea breeze further enhanced low-level moisture content, despite the already moist condition on that day.

3.2 Evolution of Convective System

Under the weak vertical wind shear condition, convection developed over southern Taiwan around 1200 LST and gradually moved northward. Upon entering the observational range of the S-Pol radar (Fig. 5a), the convective system merged with

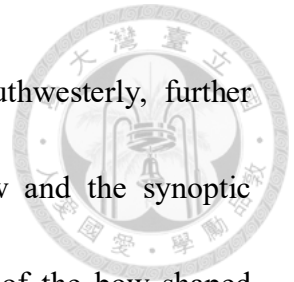


smaller convective cells that had developed over the foothills of Taichung and Miaoli by 1448 LST. By 1536 LST (Fig. 5b), a QLCS approximately 90 km in length had already formed. By 1548 LST (Fig. 5c), the radar reflectivity displayed a distinct bow-shaped structure. From 1536 to 1625 LST (Figs. 5b–d), the system moved in a southeast–northwest direction toward the coast of Taichung. By 1712 LST (Fig. 5e), the southern part of the rapidly moving convective system moved offshore and gradually dissipated, while another segment veered northward, passing over the S-Pol radar site and moving offshore by 1812 LST (Fig. 5f), followed by the development of extensive stratiform cloud cover. The entire convective system dissipated around 1912 LST (Fig. 5g).

3.3 Surface Observations

Surface observations (Fig. 6) indicated that as the convective system moved from south to north, station temperatures at stations along its path decreased over time, surface pressure (after removing semi-diurnal variations) increased, and wind direction shifted. These changes suggested the possible presence of a sustained cold pool during the development. Notably, the Qingshui station recorded the most significant temperature drop of -6 K and a surface pressure increase of approximately 2 hPa, indicating a potentially stronger cold pool as the convective system moved from inland to the coast. Additionally, at 1800 LST, as the convective system passed over the S-Pol radar site and moved offshore, observations from the Xinwu station indicated an increase in surface

wind speed and a shift in wind direction from northerly to southwesterly, further suggesting the presence of a cold pool. After the case overview and the synoptic environmental conditions, we now turn to the detailed evolution of the bow-shaped convective system.



4. Bow-shaped Echo



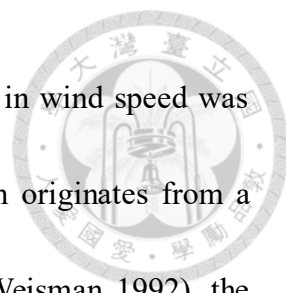
4.1 Formation and Evolution

Following the formation of the QLCS, a transient bow-shaped echo rapidly emerged. Figure 7 presents the evolution of storm-relative winds and radar composite reflectivity during this stage. Storm-relative winds were calculated by subtracting the system's leading-edge motion, which propagated northwestward at approximately 6.32 m s^{-1} between 1536 and 1637 LST. At 1536 LST (Fig. 7a), following the merger into a QLCS, a strong updraft zone developed along the leading edge, accompanied by inflow coming from the southern flank. By 1548 LST (Fig. 7b), a distinct bow-shaped echo had formed on the southern portion of the system, characterized by a banded updraft region along its leading edge. The reflectivity pattern showed that the echo center moved ahead of the northern and southern flanks, consistent with earlier findings described Wakimoto et al. (2015).

The echo curvature persisted through 1600 LST (Fig. 7c), with a pronounced updraft band along the leading edge. As the system approached the coastline by 1625 LST (Fig. 7d), updrafts weakened and the bow-shaped structure became less defined. The weakening of the southern segment may have been influenced by the presence of a marine boundary layer (Lombardo and Colle, 2012); however, the specific processes involved remain unclear and warrant further investigation. Nonetheless, some strong linear echoes

were still present along the leading edge. Concurrently, the northern segment, still over land, began a gradual northward progression. Fig. 7e-h depict the dissipation of the southern segment and the northward movement of northern segment toward the S-Pol radar, which will be discussed further in Chapter 5.

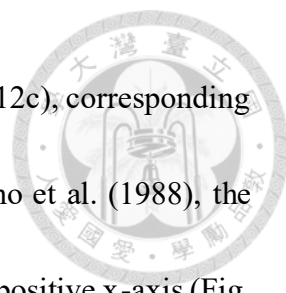
In addition to radar observations, surface station characteristics were also analyzed. Figure 8 presents temperature anomalies and accumulated precipitation during this period. Accounting for the influence of longwave and shortwave radiation, temperature anomalies were computed following the method described in Miao and Yang (2020). The temperature base state was defined as the horizontal mean of surface temperatures from stations located below 500 m elevation, spanning from Changhua to Taoyuan, during periods when convection was active. To capture evolving environmental conditions, the base state was treated as a function of time. Temperature anomalies were then derived by subtracting this time-varying base state from the observed temperatures at each station, resulting in the anomaly fields shown in Fig. 8. Figures 8a–d demonstrate that higher radar reflectivity areas corresponded with stronger accumulated rainfall. Negative temperature anomalies, associated with evaporative cooling from rain drops, were evident beneath the heaviest precipitation. Notably, the curved region produced greater rainfall and more extensive negative temperature anomalies than the linear northern portion (see pink contour in Fig. 8b).



Near the apex of the bow echo at 1548 LST, a sharp increase in wind speed was observed (Fig. 9b). Unlike the classic rear inflow jet (RIJ), which originates from a stratiform region behind the convective line (Houze et al. 1987; Weisman 1992), the current case lacked a well-defined trailing stratiform region. Although the wind speed decreased twelve minutes later, it remained elevated along the leading edge (Figs. 9c and 9d). Figure 10 shows the vertical vorticity at 2.5 km, computed using the SAMURAI wind retrieval. Before analyzing the vorticity budget terms, it is necessary to assess whether the residual term is significant. As shown in Fig. 10a, during the first four time steps (1536–1625 LST), the residual is relatively small compared to the other terms, suggesting that the analysis is reasonably reliable during this period. At 1548 LST (Fig. 10c), a clear vorticity dipole was identified at the leading edge, with the wind maximum located between the positive and negative vorticity centers. This pattern suggests that the wind enhancement may have been linked to transient mesoscale vorticity structures. However, the vorticity dipole only persisted for about 12 minutes, and the bow echo was similarly short-lived, in contrast to longer-lasting events in previous studies.

4.2 Mechanisms of Bow-shaped Echo

To better understand the formation of this vorticity dipole and the associated bow-shaped echo, a cross-sectional analysis was conducted along the region of maximum curvature (line AA' in Fig. 12). At 1536 LST, strong updrafts and intense rainfall were



evident along the leading edge, with K_{DP} values up to 2° km^{-1} (Fig. 12c), corresponding to heavy precipitation in Fig. 8a. Following the approach in Rotunno et al. (1988), the horizontal wind was rotated so that the leading edge aligned with the positive x-axis (Fig. 11a), enabling horizontal vorticity to be projected onto the vertical plane. At around 10 km altitude, positive horizontal vorticity was found between the updraft and downdraft zones, with little vorticity in low level. By 1548 LST, the system had evolved considerably (Fig. 13). K_{DP} increased to 3° km^{-1} (Fig. 12c and 13c), and the vertical velocity field showed a deep downdraft, along with a newly developed, tilted low-level updraft (Fig. 12d and 13d). This configuration may have generated negative horizontal vorticity in the lower levels (Fig. 13d), which was possibly caused by buoyancy gradients along the leading edge, consistent with the prediction from the two-dimensional horizontal vorticity equation:

$$\frac{\partial \eta}{\partial t} = -\frac{\partial B}{\partial x} \quad (1)$$

and previous findings (Rotunno et al. 1998; Weisman 1992, 1993, 1998). The new updraft led to the tilting and stretching of horizontal vorticity, forming the vertical vorticity dipole seen in Figs. 10b and 11b. This marked a clear contrast to the earlier structure observed at 1536 LST (Fig. 12d). The intensified downdraft may have led to buoyancy gradients that were possibly key for vertical vorticity dipole (Fig. 13d), which in turn could have enhanced flow curvature.

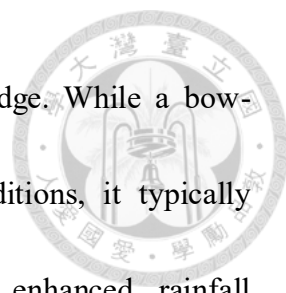


To evaluate the vorticity contribution by stretching and tilting, the vertical vorticity tendency equation from Wakimoto et al. (2015) was used:

$$\frac{d\zeta}{dt} = -(f + \zeta) \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} \right) - \left(\frac{\partial w}{\partial x} \frac{\partial v}{\partial z} - \frac{\partial w}{\partial y} \frac{\partial u}{\partial z} \right) \quad (2)$$

The first term on the right-hand side represents stretching, and the second term accounts for tilting of horizontal vorticity. Figure 10 indicates that vertical vorticity was closely aligned with regions where these two mechanisms overlapped. Using a vorticity-tendency contour of $1.0 \times 10^{-5} \text{ s}^{-2}$ and assuming a 12-minute lifespan, the estimated vorticity reaches approximately $7.2 \times 10^{-3} \text{ s}^{-1}$, which was consistent with the observed magnitude of vertical vorticity. Furthermore, during the period when the echo curvature was most evident (Figs. 10b and 10c), the spatial patterns of the tilting and stretching terms generally aligned with the vertical vorticity field, particularly in the southern portion in Fig. 10c where stretching was dominant (not shown). After 1625 LST (Fig. 10d), the correspondence weakened, indicating that other processes might have played a greater role in the later stages.

Weisman (1993) conducted numerical simulations exploring the evolution of linear convective systems under varying magnitudes of low-level vertical wind shear (VWS). When the mid-to-low-level VWS was $\leq 10 \text{ m s}^{-1}$, the simulations revealed the development of a sloped updraft and precipitation core just behind the gust front. This configuration favored a strong downdraft that reached the surface and produced



widespread, system-scale outflow extending toward the leading edge. While a bow-shaped echo could briefly emerge under these weak-shear conditions, it typically dissipated quickly. Consistent with these findings, areas of enhanced rainfall accumulation in the present case corresponded well with outward-spreading negative temperature anomalies (Fig. 8b), supporting the notion that bow-shaped echoes may transiently develop in weak-shear environments, as suggested by Weisman (1993). The region of temperature anomalies below -1 K aligned with the enhanced curvature zone (Fig. 8b), consistent with James et al. (2006), who found that stronger cold pools were often associated with more curved linear systems. As previously discussed, the vertical vorticity dipole observed in this case differed from those described in classical studies. Significantly, no prominent stratiform region developed, suggesting that the vorticity generation mechanism was unlikely related to a traditional rear-inflow jet (RIJ). Instead, observational evidence pointed to cold-pool-driven processes at the system's leading edge as the primary source of vorticity, consistent with the "subsystem vortices" described by Weisman (1998). Furthermore, a similar scenario was observed to that documented by Klimowski et al. (2004), who statistically analyzed cases over the U.S. Great Plains where isolated convective cells merged and subsequently evolved into bow echo. A comparable merger-driven development was identified in the present case (Fig. 5a–c), suggesting that quasi-linear convective system in Taiwan may exhibit hybrid characteristics. These

findings imply that the formation of bow echoes in this region is likely governed by a combination of mechanisms, rather than a single dominant process as proposed in previous studies.



Chapter 5. Movement of convection and strong surface wind

After the development of the vertical vorticity dipole and transient bow-shaped echo described in Chapter 4, the convective system continued to evolve.

5.1 Formation and Evolution

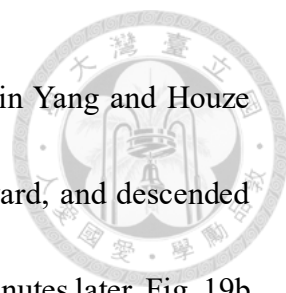
As shown in Fig. 5e-5g, the convective system passed over the S-Pol radar site and continued moving offshore. At Xinwu station, a marked increase in surface wind speed and a shift in wind direction were observed, where prevailing northerly winds veered to southwesterly. The wind profiler data from Xinwu (the red box in Fig. 14) indicated that the wind enhancement was primarily confined to the lower troposphere (below 3 km), with negligible changes aloft. After the QLCS split, the northern segment continued propagating northward. To understand the system movement and its relationship with surface wind changes, this study examined high-resolution RHI data along the 183° and 0° azimuths (Fig. 15), corresponding to the system's path between 1637 and 1912 LST. Horizontal radar reflectivity and surface observations further supported the analysis. As shown in Fig. 16, the system began to split into two convective cores (denoted A and B) around 1637 LST. Core A moved offshore and dissipated rapidly, while core B moved northward and passed over S-Pol site at approximately 1732 LST. Composite reflectivity maps showed that after passing S-Pol, the northern system was followed by a stratiform region, which developed a well-defined bright band after 1700 LST (Fig. 16d). This bright

band can be further confirmed by the vertical structure seen in the RHI radar reflectivity cross-sections shown in Figs. 24. Corresponding temperature anomaly maps (Fig. 17) and time series plots (Fig. 6) indicate that this northward movement was accompanied by surface cooling and heavy rainfall, suggesting that cold pool processes remained active.

5.2 Rear-to-Front Flow and Bounded Weak Echo Region

Figure 18b shows that the radar reflectivity and K_{DP} columns consisted of multiple convective cells at various stages of development. New cells were forming near the leading edge, approximately 20 km upstream of S-Pol radar. At this time, the 35-dBZ echo top ranged from 6 to 10 km in height. Figure 18c reveals a sloping updraft ahead of the leading edge and enhanced low-level downdrafts. Temperature fields in Fig. 8e suggest the presence of two cold pools: one from the earlier southward-moving system and the other from the relatively stationary northward system. A horizontal temperature gradient pointed toward the S-Pol indicates that surface downdrafts had strengthened the cold pool, enhancing low-level radial inflow.

In the trailing region of the system, a midlevel flow was observed descending from approximately 4–5 km altitude in the radial velocity field (Fig. 18c), supported by the environment sounding data. This flow originated in a dry midlevel layer and transported cool, dry air into the rear of the system. This pattern is consistent with the descending rear-inflow described by Houze et al. (1987) and Weisman (1992). A similar rear-to-front



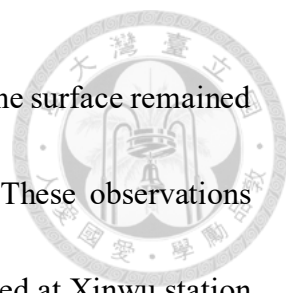
(RTF) flow was later observed, resembling the structure described in Yang and Houze (1995). This RTF flow started above the melting layer, moved forward, and descended near the surface approximately 40 km south of S-Pol radar. Twelve minutes later, Fig. 19b shows the continued development of new convective cells ahead of the system, which merged into the main storm and maintained northward propagation. The echo top reached approximately 12 km. Figure 19c shows a cold pool signature within the 0–2 km layer, located approximately 20 km from the S-Pol radar. This feature is supported by surface observations in Fig. 8f, where negative temperature anomalies are evident near the leading edge of the system. The leading updraft tilted upwind, and the rearward flow extended farther forward compared to that in Fig. 18c. The RTF flow continues to propagate forward, as shown in Fig. 20c. At this time, the maximum vertical velocity (w) exceeds 20 m s^{-1} . According to the hypothesis proposed by Houze (1987) and supported by numerical simulations conducted by Yang and Houze (1995), the mesoscale low-pressure perturbations associated with the stratiform and convective regions interact synergistically, establishing a continuous RTF flow throughout the entire system. As shown in Fig. 19c, the cold pool's negative buoyancy became dominant, leading to a more inclined updraft and possible generation of midlevel negative pressure perturbations in convective region. Evaporative cooling within the stratiform region might further strengthen another midlevel low-pressure perturbation, allowing the rear inflow to intrude

toward the leading convective region.

At 1708 LST, the dark indigo box in Fig. 21b shows the formation of a single K_{DP} column, with its top extending above the 0°C isotherm, which was estimated at approximately 4–5 km altitude based on the sounding-derived melting level. Since the depth and volume of the K_{DP} column are generally positively correlated with the strength and size of updrafts (van Lier-Walqui et al. 2016), suggesting that the system exhibited stronger updrafts at this time than earlier. Figure 21b also showed that A weak reflectivity zone appeared at mid-to-lower levels along the leading edge, partially surrounded by a hook-shaped region of higher radar reflectivity. The corresponding decrease in ZDR (the dark indigo box in Fig. 21d) implies smaller raindrop sizes in this region. This feature was consistent with the bounded weak-echo region (BWER) described by Chisholm (1973), which commonly occurs in convective systems with intense updrafts. The strong vertical updrafts (Fig. 21c) in the lower levels at the storm's leading edge could rapidly lift small hydrometeors upward before they could have sufficient time to grow, resulting in a localized minimum in radar reflectivity.

5.3 Surface wind intensification

At 1732 LST, Fig. 22 indicates that the RTF flow had reached the cold pool and arrived at the leading edge of the convective storm. The previous RHI scan from the S-Pol radar showed near-surface radial wind speeds exceeding 21 m s⁻¹ up to 4-km altitude.



After the system passed the S-Pol radar, outbound wind speeds near the surface remained around 17 m s^{-1} , and low-level precipitation persisted (Fig. 17c). These observations suggest that the strengthening and turning of low-level winds observed at Xinwu station were likely due to a combination of advancing RTF flow and outflow from precipitation evaporation. However, since the system was already weakening (Figs. 22–24), the surface wind enhancement was moderate and did not reach damaging levels (Weisman 1992; Mahoney et al. 2011).

5.4 System Dissipation

To examine the dissipation process of this convective system, we apply the Rotunno–Klemp–Weisman (RKW) theory (Rotunno et al. 1988), following the approach of Pan and Yang (2022) and Mashiko (2024). According to the RKW theory, under hydrostatic conditions, when the horizontal vorticity generated by the cold pool exceeds that produced by the low-level environmental wind shear, a linear convective system tilts upshear, resulting in weaker lifting and eventual dissipation. As the system approached the Xinwu station, wind profiler data were used to calculate the vertical wind shear perpendicular to the system’s motion. From 1640 to 1800 LST, prior to the wind shift, the low-to-mid-level wind shear was approximately 2 to 4 m s^{-1} (Fig. 25). Assuming hydrostatic balance and neglecting surface friction, we estimated the cold pool intensity by adapting the formulation proposed by Benjamin (1968; Eq. 1.1):

$$c = \sqrt{2gH \frac{\rho - \rho_0}{\rho_0}} \text{ (Benjamin, 1968)}$$

$$\sim \sqrt{\frac{2(P - P_0)}{\rho_c}} \text{ where } \rho_c \approx 1 \text{ kg/m}^3$$

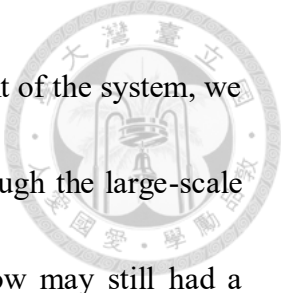


Based on Fig. 6, which shows a ~2 hPa surface pressure difference at the Hsinchu station during the period of maximum surface cooling, we derived an estimate of the cold pool strength. Assuming negligible surface friction, the estimated cold pool intensity was approximately 20 m s⁻¹, which far exceeded the aforementioned environmental vertical wind shear. A simple estimate of the gust front propagation speed at the leading edge of the system, based on radial velocity observations, yielded a value of about 11 m s⁻¹. This speed was reasonably lower than the cold pool velocity under the frictionless condition, yet still significantly higher than the environmental shear. These findings suggest that the system's tilt had been dominated by cold-pool dynamics, causing the storm to tilt upshear and rapid weakening.

There were several possible reasons for the weakening of the updraft and convection. For example, tilted updrafts may experience greater mixing with surrounding environment air, which can reduce buoyancy and result in a weaker convective system compared to one with more upright updrafts (Weisman 1992; Mulholland et al. 2021). Upright convective systems tended to gain more positive buoyancy, as the downward-directed perturbation pressure gradient force caused by vertical buoyancy gradients was

weaker than that in tilted systems (Parker 2010). In addition, Mashiko (2024), through numerical simulations, discussed that in the context of QLCS dissipation, the weakening of updrafts might result from the vertical decrease of the pressure perturbation gradient term in the presence of a strong cold pool, ultimately leading to the decay of the convective system. This aspect warrants further investigation in future studies.

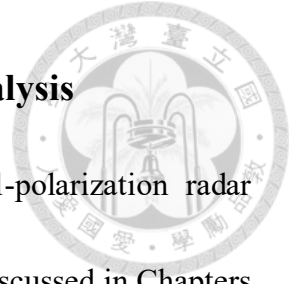
Regarding the northward propagation of the convective system, we hypothesize that it was closely related to the location of newly generated convective cells. Based on temperature anomalies (Figs. 8 and 17) and surface wind observations (Fig. 26), new cells likely formed due to cold-pool expansion and interaction with the sea breeze under weak synoptic conditions. This supports the idea that cold pool dynamics played a key role in initiating new convection. As the cold pool expanded outward and interacted with the sea breeze, low-level convergence intensified along the leading edge. Given the generally moist surface conditions in Taiwan during summer, areas with greater thermodynamic instability were more favorable for convective development. Xie and Scofield (1989) noted that in the absence of strong synoptic forcing, convection tended to initiate and propagate toward regions with higher equivalent potential temperature (θ_e) and θ_e ridges. Surface station data showed a θ_e maximum near the S-Pol radar site (Fig. 27), suggesting the system had preferred development in that direction. Given the typically moist lower atmosphere, the reduction in moisture likely resulted from the descent of dry



midlevel air within the cold pool. To further understand the movement of the system, we also consider the influence of the synoptic-scale environment. Although the large-scale forcing was relatively weak during this period, the background flow may still had a broader influence. Sounding data from Hsinchu (Fig. 28a and 28b) showed that low-level winds lacked a consistent directional trend and were sometimes northwesterly. In contrast, the Makung sounding (Fig. 28c) indicated predominantly southwesterly winds near the surface. The 925-hPa wind field (Fig. 28d) also showed southwesterly flow in the vicinity of Hsinchu, suggesting that synoptic-scale winds may have supported the system's northward movement. However, without the continuous development of new convective cells, the system might not have maintained its northward propagation and would have dissipated earlier. Therefore, both synoptic-scale and mesoscale processes likely contributed to the system's evolution. The topography of Taiwan may also have influenced the system's movement, and this aspect should be explored further in future studies. Cold pool passage was further supported by MPD observations (Fig. 4). Before 1732 LST, the low-level mixing ratio between 250 and 600 m was around 24 g kg^{-1} . After the cold pool passed, the vapor mixing ratio decreased to about $18\text{--}20 \text{ g kg}^{-1}$. Given the typically moist lower atmosphere, the reduction in moisture likely resulted from the descent of dry midlevel air within the cold pool.



Chapter 6. Dual-Polarization Parameter Analysis

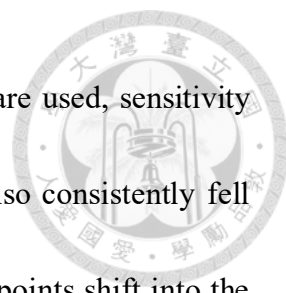


This chapter examines the statistical characteristics of dual-polarization radar parameters to further investigate the microphysical interpretations discussed in Chapters 4 and 5. In particular, we examine the evolution of warm-rain processes and related radar signatures based on CFADs (Contoured Frequency by Altitude Diagrams), using both the PPI and RHI cross sections.

6.1 Z_H - Z_{DR} Difference Space

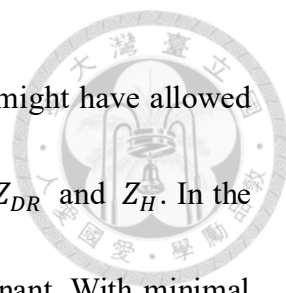
To categorize the dominant microphysical processes, we have adopted the diagnostic parameter space proposed by Kumjian et al. (2014, 2022), in which the horizontal axis shows the vertical difference in reflectivity (Z_H) and the vertical axis shows the difference in differential reflectivity (Z_{DR}) between the surface and the bottom of the melting layer. This parameter space allows for the classification of warm-rain processes into four quadrants, each corresponding to different microphysical mechanisms (see Fig. 4 of Kumjian 2022). Based on RHI and sounding data, the melting layer for this case study was located at approximately 4–5 km altitude; therefore, the warm-rain region was defined between 1.5 and 3 km above sea level. Grid-based analysis was applied to the S-Pol PPI radar data within a 30–100 km radial annulus (Fig. 29a).

Figure 29b displays the temporal evolution of averaged values. All analyzed time periods fall within the second quadrant, typically associated with microphysical processes



dominated by size sorting and evaporation. Although mean values are used, sensitivity tests were conducted using the 50th and 90th percentiles, which also consistently fell within the second quadrant. Only at the 99th percentile did all data points shift into the first quadrant. Therefore, the mean is considered a more representative choice for characterizing the overall microphysical state during this period. We first examine the three early observation periods (Fig. 29; blue line), which correspond to the formation and peak stage of the bow-shaped echo. As described in Chapter 4, this phase featured strong low-level reflectivity and significant evaporative cooling, both of which contributed to cold-pool formation and radar echo curvature. From the perspective of dual-polarization radar characteristics, these time periods show an increase in Z_{DR} toward lower levels within the warm-rain layer, accompanied by a decrease in Z_H . Two microphysical mechanisms could explain this observed pattern. The first is size sorting in a vigorous convective environment, where larger drops fall more rapidly while smaller drops remain aloft, producing a vertical separation that increases Z_{DR} and reduces Z_H . The second possible explanation is related to evaporative cooling, which tends to remove smaller drops. This process reduces the total drop concentration, leading to a decrease in Z_H , while the relative dominance of remaining large drops enhances Z_{DR} .

From 1625 to 1748 LST (Fig. 29; green line), as the main convective region moved offshore and gradually weakened, the bow-shaped structure became less pronounced.



Reduced rainfall and weaker evaporative cooling during this stage might have allowed smaller droplets to persist, resulting in weaker vertical gradients in Z_{DR} and Z_H . In the final stage (Fig. 29; red line), stratiform precipitation became dominant. With minimal vertical wind shear and weak dynamic forcing, hydrometeors experienced little external influence during descent. Larger drops settled more quickly, while smaller ones were gradually removed by evaporation. Dry air also began to extend into the warm-rain layer from the rear of the system, further reducing the number concentration of small drops. As a result, microphysical parameters remained concentrated in the second quadrant. These findings suggest that although similar dual-polarization radar signatures were observed, their underlying microphysical causes differed across precipitation types.

6.2 RHI CFADs during the Bow-shaped Echo

Due to the inherent limitations of gridded PPI data, especially at near and far ranges from the radar, we further analyzed the high-resolution RHI dataset collected by S-Pol. Following the method of McMurdie et al. (2018), who applied the RHIs for CFAD (Contoured Frequency by Altitude Diagrams) analysis first proposed by Yuter and Houze (1995), we adapted a similar framework to investigate the microphysical processes associated with this convective system. Although RHIs offer limited horizontal coverage, we selected cross sections intersecting key convective structures and grouped them into 10-km radial intervals. Within each radial interval, median values of microphysical

parameters were computed across azimuths to enable detailed comparisons and elaborate physical interpretations made in earlier chapters.



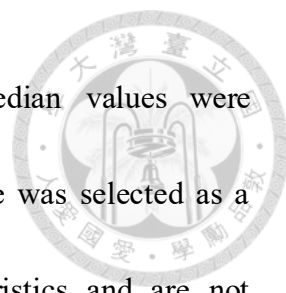
In Chapter 4, we focused on the development of the bow-shaped echo, particularly at 1536 and 1548 LST. These two time periods represent the initial formation and peak intensity of the bow structure. RHI scans between 195° and 205° were analyzed, with Fig. 31 showing the vertical reflectivity structure. At both times, low-level reflectivity between 60 and 80 km exceeded 40–50 dBZ with steep vertical gradients, indicating strong convective activity. At 1548 LST, the vertical extent of high reflectivity increased, implying more vigorous updrafts and stronger convection. A closer examination of Fig. 31e to 31g reveals that the bow-shaped echo features can be identified from three different viewing angles. The 205° RHI, located at the leading edge of the bow, shows the strongest low-level reflectivity at the bending region (Fig. 31f). In contrast, Figure 31e and Fig. 31g (60–70 km and 80–90 km) indicate lower or absent reflectivity due to the rearward displacement of the echo flanks caused by the curvature of the system. The 200° RHI was located slightly rearward of the bow apex but still within the high-reflectivity region; thus, low-level reflectivity did not vary significantly with distance. On the other hand, the 195° RHI, situated on the rear side of the bow-shaped echo, shows the weakest low-level reflectivity (80–90 km range), with a slight increase at 90–100 km and 60–80 km, consistent with features expected on the backside of a bow echo. Figure 32 presents the



vertical profiles of K_{DP} , which show relatively large values of K_{DP} in the 60–80 km range at low levels, indicating higher liquid water content and the potential for heavy rainfall. In particular, the 200° cross section (Fig. 32b), located near the bow-shaped echo, shows significantly enhanced K_{DP} values compared to Fig. 32f, with most of the strong signal concentrated near the bend. Other radial directions exhibit much weaker signals, consistent with earlier surface observations and physical descriptions. This parameter (K_{DP}) also captures the bow echo features across different azimuths. For instance, in the 205° RHI, slightly higher K_{DP} values (up to $0.7^\circ \text{ km}^{-1}$) were observed at 70–80 km, while values before and after the curvature remain near 0° km^{-1} . In contrast, at 195° RHI, stronger K_{DP} appeared on both flanks of the bow-shaped echo, while the region behind the bend, where echo strength was weaker, showing lower K_{DP} values. Although median raindrop size remains relatively stable (Fig. 33), Z_{DR} at 1548 LST decreases with height (Fig. 33f and Fig. 33g), suggesting a reduced concentration of small drops near the surface. Combined with the earlier inference of strong rainfall and downdrafts, this pattern supported the presence of enhanced evaporation, consistent with size sorting and downdraft-driven processes, which might have contributed to more prominent bow-shaped echo structure.

6.3 RHI CFADs in the Stratiform region

For the stratiform region, a different analytical approach was applied due to its broad



spatial extent and more uniform precipitation distribution. Median values were azimuthally averaged over the stratiform area. The 1812 LST case was selected as a representative time, as other periods exhibited similar characteristics and are not discussed here in detail. A well-defined bright band near 4–5 km altitude was evident across multiple radar variables, including reflectivity, differential reflectivity, and cross-correlation coefficient (Fig. 34). This bright band resulted from melting ice particles, which retained a solid outer layer while their exterior transitioned to liquid, enhancing the dielectric constant and thus radar reflectivity in the melting layer. Fig. 35 compares the median vertical reflectivity profiles between convective and stratiform time periods. Similar differences were evident in dual-polarization parameters as well. In the stratiform region, CFADs show highly consistent vertical distributions across azimuths, with narrow spreads indicating spatial uniformity (Fig. 35b). In contrast, the convective CFADs displayed broader distributions, reflecting greater variability consistent with the heterogeneous nature of convective processes (Fig. 35a).

6.4 RHI CFADs of Rear Inflow

Chapter 5 focuses on the evolution of the rear inflow, which progressively extended forward and downward from middle levels, transporting dry air from the rear of the storm into the lower troposphere. This inflow enhanced evaporative cooling of precipitation and intensified surface winds. To assess the influence of downdrafts near the leading edge of



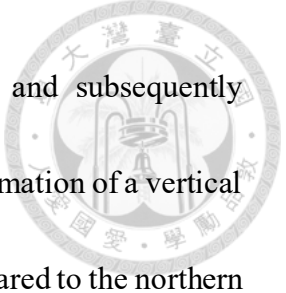
convection and the intrusion of dry air, we examined the 30–40 km range in the RHI domain, corresponding to the region where dry air was detected near 40 km from the radar. This analysis focuses on RHIs between approximately 185° and 201°, as highlighted in Fig. 36. In the layer below 4 km, the radial wind component within this range shows an increase in RTF flow after 1700 LST, with initially weak negative values becoming more intense (Fig. 36a and 36b). In the warm-rain region, the median reflectivity initially shows minimal vertical variation. However, after 1700 LST, as dry air advanced into the 30–40 km range, reflectivity began to decrease with height even in the lower levels (Fig. 36g to 36i). A similar pattern was observed in Z_{DR} , which transitioned from a vertically uniform profile to one that decreased with altitude (Fig. 36f). This change was consistent with enhanced evaporation caused by descending dry air. It is hypothesized that smaller raindrops were preferentially evaporated, leading to an overall reduction in hydrometeor number concentration and, consequently, a decrease in radar reflectivity. At the same time, the relative abundance of larger drops increased, resulting in higher Z_{DR} values that produced a signature commonly linked to warm-rain processes. Therefore, the temporal evolution of dual-polarization radar variables in RHIs provided observational evidence for the gradual forward progression of the RTF flow.

Chapter 7. Summary



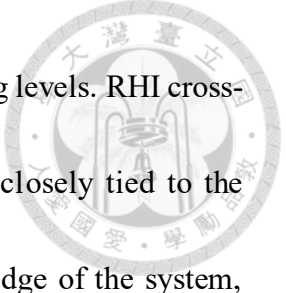
In this thesis study, observations from the NCAR S-Pol radar and enhanced sounding profiles collected during the TAHOPE field campaign are used to analyze a quasi-linear convective system (QLCS) that occurred on 29 June 2022 (SOP4) over the western plains of Taiwan. On the day of this event, no prominent synoptic-scale system was present near Taiwan. The environment featured weak low-level vertical wind shear but relatively high CAPE. During the evolution of the convective system, a short-lived bow-shaped echo developed over the Taichung to Miaoli plains, accompanied by the emergence of vorticity couplets. After the system split, the northern segment continued propagating northward and passed over the S-Pol radar site. Around this time, the Xinwu surface station recorded a sudden shift in wind direction and an increase in wind speed. Shortly thereafter, the convective system rapidly dissipated. Throughout the merging and dissipation phases of the QLCS, cold pools were continuously generated at the surface.

The mechanisms underlying the transient development of the bow-shaped echo are illustrated in the conceptual diagram shown in Fig. 37. As precipitation intensified, the associated downdrafts became more pronounced in the lower troposphere. Evaporative cooling within these downdrafts led to the formation of a cold pool, which, created a horizontal buoyancy gradient. This buoyancy gradient promoted the generation of cold-pool-induced horizontal vortices. The outflow boundary of the cold pool contributed to



the curvature of the convective echo. The upward motion tilted and subsequently stretched the system's inherent horizontal vorticity, resulting in the formation of a vertical vorticity couplet near the location of maximum echo curvature. Compared to the northern and southern flanks, the apex of the bow-shaped echo exhibited greater accumulated rainfall, indicating stronger instantaneous rainfall intensity in that region. As the system evolved, the curved convective structure could not be maintained due to the weak environmental vertical wind shear and its progression toward the more stable marine boundary layer. Consequently, the bow-shaped echo rapidly weakened and dissipated.

Several notable dynamical features were observed during the northward progression of the convective system, which splits from the main QLCS after part of the system moved offshore. These processes are summarized in the conceptual diagram shown in Fig. 38. As the system gradually propagated northward, a rear inflow developed and extended forward with time. This rear-to-front flow (RTF) likely resulted from interactions between perturbation pressure fields associated with the stratiform and convective regions. After 1700 LST, the system approached the S-Pol radar site, and the advancing RTF flow penetrated into the leading convective edge. Its interaction with the precipitation-induced outflow led to a wind shift and intensification at the nearby Xinwu surface station. However, based on the RKW theory, the cold pool at this time had become substantially stronger than the ambient vertical shear, indicating that the system had entered a



dissipative phase. As a result, the surface winds did not reach damaging levels. RHI cross-section analyses further suggest that the motion of the QLCS was closely tied to the initiation and development of new convective cells. At the leading edge of the system, convergence between the cold pool outflow and the sea breeze front provided a favorable environment for convective initiation. In the absence of a synoptic-scale forcing, the system also tended to propagate toward regions of higher equivalent potential temperature (θ_e), particularly near the S-Pol location, where greater instability likely contributed to continued convective development.

Although this study has utilized multiple observation datasets to analyze the physical mechanisms governing the development and dissipation of the QLCS during SOP4, limitations remain due to the discontinuous nature of the observations. As a result, it is difficult to fully address the continuous evolution of the associated physical processes. Moreover, the current analysis primarily focuses on dynamical aspects, while the thermodynamic aspect is relatively less discussed, relying mainly on surface station data and occasional sounding data. To further investigate the three-dimensional thermodynamic structure and better understand the roles of storm motion and cold-pool dynamics, future work could incorporate thermodynamic retrieval techniques or employ convection-permitting numerical simulations. These approaches would allow for a more continuous depiction of convective evolution and enable detailed examination of the

underlying physical mechanisms. Such efforts are essential to gain a more comprehensive understanding of the Q LCS behavior in weakly forced summer environments in Taiwan.



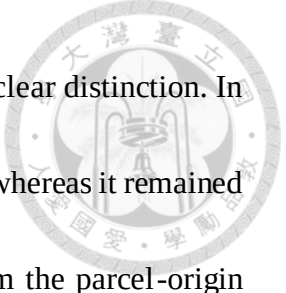
Appendix A
Environmental Parameter Differences
Between CI and nonCI Events in TAHOPE



Previous studies have shown that favorable environmental conditions for convective initiation (CI) vary by region and are influenced by different factors. We use the sounding data from the Taiwan-Area Heavy Rain Observation and Prediction Experiment (TAHOPE) to examine environmental differences between CI and non-CI days, with specific parameters and methods following those described by Marquis et al. (2023).

We selected the Hsinchu radiosonde data as the analysis target because it provided the 06 UTC sounding even on non-CI days. Under weak synoptic condition, CI was defined based on radar criteria: composite reflectivity exceeding 20 dBZ for at least 30 minutes over 5 km², and reaching 35 dBZ for at least 20 minutes. Days with weak synoptic conditions but no rainfall were categorized as non-CI. Cases affected by Mei-Yu fronts, typhoons, or strong background winds were excluded. The cases that featured additional soundings and fell under weak synoptic conditions are listed in Table A.

Figure A1 to A3 showed that mid-level relative humidity (500–700 hPa) was notably higher in CI cases (maintaining above 60%) than in nonCI cases (often below 60%). Wind profiles showed weak shear and slight southwesterly flow in both cases. CI cases also had higher cooling rates in the lower levels. While the surface CIN often dropped to near zero



by midday in both scenarios, mixed-layer CIN (MLCIN) provided a clear distinction. In CI cases, MLCIN typically dropped below -10 J/kg before initiation, whereas it remained above -20 J/kg or even increased in nonCI cases. The distance from the parcel-origin level to the ML LFC ($ML \Delta z$) was also shorter in CI cases, with higher lapse rates (from parcel-origin level to ML LFC), reducing dilution and promoting successful initiation. Topography played a key role. Most CI events occurred on gentle slopes (200–500 m) rather than on flat plains on Taiwan. Weak background winds and surface heating, combined with terrain-induced upslope flows, helped lift parcels to the LFC. In summary, the overall differences observed in the selected cases during TAHOPE were primarily influenced by the dryness of the synoptic-scale environment, with boundary layer thermodynamic conditions playing a secondary role.

This thesis study also conducted a composite analysis of other CI and non-CI cases that met the selection criteria to examine the prevailing synoptic weather conditions. The findings were broadly consistent with previous research (e.g., Lin et al., 2011). Additionally, by using the MPD (Fig. A4) and selecting the lowest stable altitude with available data (521 m), it was found that both CI and non-CI cases exhibited signs of moisture being advected inland by the land-sea breeze. However, in non-CI cases, this

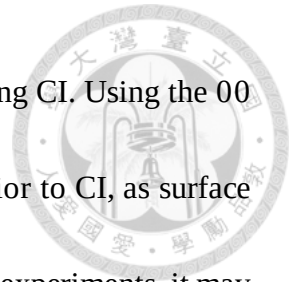
inland moisture transport weakened after 11–12 LST and did not persist, whereas in CI cases, higher moisture amounts were maintained during this period.



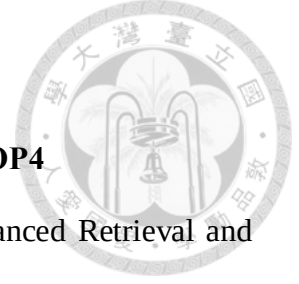
Although this study has yielded some results, it still faces certain limitations, including representativeness issues and other challenges. The initial development regions of CI events can be categorized into foothill areas at elevations of 500–1000 m or mountaintops. Whether soundings and surface observations from plains regions can adequately represent the pre-CI environment remains debatable. Marquis et al. (2023), in their observational study near the Argentine mountains, ultimately suggested that the temporal resolution of sounding observations should be shorter than 1 hour and the spatial resolution should be less than $O(10 \text{ km})$. However, in TAHOPE, logistical and budget constraints prevented the release of soundings in mountainous areas, and the spatial density of observation stations in mountainous or foothill regions was sparse. Given the heterogeneous distribution of land cover in mountainous areas, boundary layer conditions there might differ significantly from those on the plains. Thus, the results of the analysis presented in this Appendix may not be fully applicable to convective initiation over mountainous regions.

Furthermore, although routine soundings were launched at 06 UTC at Hsinchu station during TAHOPE, in most CI events, vigorous convection had already developed by that

time, potentially obscuring the environmental characteristics preceding CI. Using the 00 UTC sounding alone might be too early to capture the conditions prior to CI, as surface heating might not yet be fully established. For future large-scale field experiments, it may be advisable to adjust the standard sounding schedule (00, 06, 12, 18 UTC) to include (or add) 03 and 04 UTC soundings, which would be more beneficial for capturing the actual thermodynamic and dynamic characteristics of the pre-convection environment. Additionally, if operationally feasible, releasing a series of soundings along a transect from the plains into the mountains may provide further insights into boundary layer differences between plains and mountainous regions in Taiwan.



Appendix B
Spline Analysis at Mesoscale Utilizing Radar
and Aircraft Instrumentation (SAMURAI) for SOP4



The SAMURAI (Spherical and Cartesian Mapping for Unbalanced Retrieval and Analysis of Integrated Observations) system is a variational analysis framework that yields a maximum likelihood estimate of the atmospheric state by minimizing a cost function based on a given set of observations and associated error statistics.

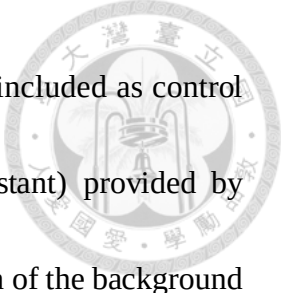
The variational cost function and its gradient, following Bell et al. (2012), are given by:

$$J(\hat{\mathbf{x}}) = \frac{1}{2} \hat{\mathbf{x}}^T \hat{\mathbf{x}} + \frac{1}{2} (\mathbf{H}\mathbf{C}\hat{\mathbf{x}} - \mathbf{d})^T \mathbf{R}^{-1} (\mathbf{H}\mathbf{C}\hat{\mathbf{x}} - \mathbf{d}) \quad (B1)$$

$$\nabla J(\hat{\mathbf{x}}) = (\mathbf{I} + \mathbf{C}^T \mathbf{H}^T \mathbf{R}^{-1} \mathbf{H}\mathbf{C}) \hat{\mathbf{x}} - \mathbf{C}^T \mathbf{H}^T \mathbf{R}^{-1} \mathbf{d} \quad (B2)$$

Here, $\mathbf{H}$ is the linearized observation operator, $\mathbf{C}$ is the square root of the background error covariance, $\mathbf{R}$ is the observation error covariance and $\mathbf{d}$ is the observation increment. The cost function in Equation (B1) is minimized using a conjugate gradient algorithm, with the initial condition $\hat{\mathbf{x}} = \mathbf{0}$.

The boundary condition follows the method proposed by Ooyama (2002). In this study, the R1T1 option is applied, which imposes the condition $u' = 0$. The lower boundary condition is specified using an immersed boundary method (IBM) to incorporate Taiwan's complex topography. The SAMURAI analysis is configured to retrieve three-dimensional wind components (u , v , w) on a Cartesian grid.



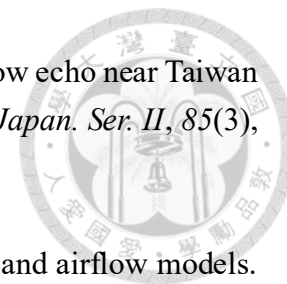
Thermodynamic variables such as temperature and pressure are not included as control variables and are instead held at default background values (constant) provided by SAMURAI. Since there is no background state, the standard deviation of the background error is purposefully set to 1.0 to reflect the lack of prior knowledge about the background state. This allows the analysis to rely primarily on observations. However, to mitigate potential instability in data-sparse regions and base line problem, a Gaussian recursive filter is applied during the analysis, helping to smooth the analysis, ensure spatial consistency and to suppress spurious small-scale noise.

SAMURAI employs a finite element method with cubic B-splines as basis functions. This approach offers several advantages. It allows flexible treatment of irregular boundaries (such as immersed boundary method), accommodates unevenly distributed observations, and directly assimilates radar data in its native polar coordinate system without the need for interpolation to a Cartesian grid. In addition, physical constraints such as mass continuity and momentum conservation can be incorporated directly into the variational cost function, which enhances the physical consistency of the retrieved wind and thermodynamic fields (Cha et al. 2023).

Reference

- 陳泰然、周鴻祺、紀水上、黃心怡及楊進賢. (2012). 台灣與其他地區暖季弓形回波之特徵與環境條件. *大氣科學*, **40(1)**, 49-75. [Chen, George Tai-Jen, Hon-Chi Chou, Shui-Shang Chi, Shin-Yi Huang, and Jinn-Shang Yang. "Characteristics and Environmental Conditions of Bow Echoes Occurred Over Taiwan and Other Areas." *Atmospheric Science* 40, no. 1 (2012): 49–75.]
- 陳泰然、陳來發、周蘭芬 李育祺. (1988). 1985 年與 1986 年暖季西太平洋高層冷心低壓之初步分析，氣象學報，第三十四卷，**275-284**. [A Preliminary Study on the Western Pacific Upper-Tropospheric Cold Vortices in the Warm Months of 1985-1986]
- 吳宗堯. (1976). 夏季東亞高層冷心低壓之分析. *大氣科學*, **3(1)**, 1-11. [An Observational Stz of the Cold Core Low in the upper Troposphere in Far East in Summer.]
- 楊明仁，鄭明典，黃清勇，林沛練，林博雄，李清勝，王重傑，劉清煌，呂國臣，張保亮，黃椿喜，侯昭平，鍾高陞，張偉裕，林品芳，2023：臺灣區域豪雨觀測暨預報實驗(TAHOPE)簡介，*中華民國氣象學會會刊*，**64**，50–67。
- Agee, E., and E. Jones, 2009: Proposed Conceptual Taxonomy for Proper Identification and Classification of Tornado Events. *Weather and Forecasting*, 24(2), 609-617, <https://doi.org/10.1175/2008waf2222163.1>
- Bell, M. M., M. T. Montgomery, and K. A. Emanuel, 2012: Air-Sea Enthalpy and Momentum Exchange at Major Hurricane Wind Speeds Observed during CBLAST. *Journal of the Atmospheric Sciences*, 69(11), 3197-3222. <https://doi.org/10.1175/Jas-D-11-0276.1>
- Benjamin, T. B., 1968: Gravity currents and related phenomena. *Journal of fluid mechanics*, 31(2), 209-248, [https://doi.org/ doi:10.1017/S0022112068000133](https://doi.org/doi:10.1017/S0022112068000133)
- Cha, T. Y., and M. M. Bell, 2023: Three-Dimensional Variational Multi-Doppler Wind Retrieval over Complex Terrain. *Journal of Atmospheric and Oceanic Technology*, 40(11), 1381-1405, <https://doi.org/10.1175/Jtech-D-23-0019.1>

Chen, G. T.-J., C.-C. Wang, and H.-C. Chou, 2007: Case study of a bow echo near Taiwan during wintertime. *Journal of the Meteorological Society of Japan. Ser. II*, 85(3), 233-253, <https://doi.org/10.2151/jmsj.85.233>



Chisholm, A. J., 1973: Alberta hailstorms Part I: Radar case studies and airflow models. In *Alberta Hailstorms* (pp. 1-36). Springer.

Coniglio, M. C., D. J. Stensrud, and M. B. Richman, 2004: An observational study of derecho-producing convective systems. *Weather and Forecasting*, 19(2), 320-337, [https://doi.org/10.1175/1520-0434\(2004\)019<0320:AOSODC>2.0.CO;2](https://doi.org/10.1175/1520-0434(2004)019<0320:AOSODC>2.0.CO;2)

Evans, J. S., and C. A. Doswell III, 2001: Examination of derecho environments using proximity soundings. *Weather and Forecasting*, 16(3), 329-342. [https://doi.org/10.1175/1520-0434\(2001\)016<0329:EODEUP>2.0.CO;2](https://doi.org/10.1175/1520-0434(2001)016<0329:EODEUP>2.0.CO;2)

Fujita, T. T. 1978: *Manual of downburst identification for project NIMROD*.

Grim, J. A., R. M. Rauber, G. M. McFarquhar, B. F. Jewett, and D. P. Jorgensen, 2009: Development and Forcing of the Rear Inflow Jet in a Rapidly Developing and Decaying Squall Line during BAMEX. *Monthly Weather Review*, 137(4), 1206-1229, <https://doi.org/10.1175/2008mwr2503.1>

James, R. P., P. M. Markowski, and J. M. Fritsch, 2006: Bow echo sensitivity to ambient moisture and cold pool strength. *Monthly Weather Review*, 134(3), 950-964, <https://doi.org/10.1175/Mwr3109.1>

Klimowski, B. A., M. R. Hjelmfelt, and M. J. Bunkers, 2004: Radar observations of the early evolution of bow echoes. *Weather and Forecasting*, 19(4), 727-734, [https://doi.org/10.1175/1520-0434\(2004\)019<0727:Rootee>2.0.Co;2](https://doi.org/10.1175/1520-0434(2004)019<0727:Rootee>2.0.Co;2)

Kumjian, M. R., O. P. Prat, K. J. Reimel, M. van Lier-Walqui, and H. C. Morrison, 2022: Dual-Polarization Radar Fingerprints of Precipitation Physics: A Review. *Remote Sensing*, 14(15), <https://doi.org/10.3390/rs14153706>

Kumjian, M. R., S. A. Rutledge, R. M. Rasmussen, P. C. Kennedy, and M. Dixon, 2014: High-Resolution Polarimetric Radar Observations of Snow-Generating Cells. *Journal of Applied Meteorology and Climatology*, 53(6), 1636-1658, <https://doi.org/10.1175/Jamc-D-13-0312.1>

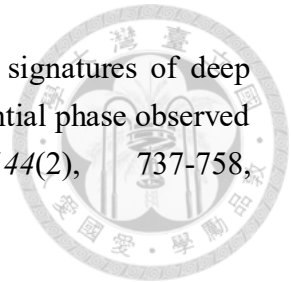


- Lin, P.-F., P.-L. Chang, B. J.-D. Jou, J. W. Wilson, and R. D. Roberts, 2011: Warm season afternoon thunderstorm characteristics under weak synoptic-scale forcing over Taiwan Island. *Weather and Forecasting*, 26(1), 44-60, <https://doi.org/10.1175/2010WAF2222386.1>
- Liu, Q., X. Xu, K. Zhao, and A. Zhou, 2023: A Merger-Formation Bow Echo Caused by Low-Level Mesovortex in South China. *Journal of Geophysical Research-Atmospheres*, 128(8), <https://doi.org/10.1029/2022JD037954>
- Lombardo, K. A. and B. A. Colle, 2012: Ambient conditions associated with the maintenance and decay of quasi-linear convective systems crossing the northeastern U.S. coast. *Mon. Wea. Rev.*, **140**, 3805–3819, <https://doi.org/10.1175/MWR-D-12-00050>
- Mahoney, K. M., G. M. Lackmann, and M. D. Parker, 2009: The role of momentum transport in the motion of a quasi-idealized mesoscale convective system. *Mon. Wea. Rev.*, **137**, 3316–3338, <https://doi.org/10.1175/2009MWR2895.1>
- Marquis, J. N., Z. Feng, A. C. Varble, T. C. Nelson, A. Houston, J. M. Peters, J. P. Mulholland, and J. Hardin, 2023: Near-Cloud Atmospheric Ingredients for Deep Convection Initiation. *Monthly Weather Review*, 151(5), 1247-1267, <https://doi.org/10.1175/Mwr-D-22-0243.1>
- Mashiko, W. 2024: Structure and Maintenance of a Quasi-Linear Mesoscale Convective System That Caused Torrential Rain in Southern Kyushu, Japan, on 11 July 2021. *Monthly Weather Review*, 152(8), 1903-1921, <https://doi.org/10.1175/MWR-D-23-0209.1>
- Mathias, L., V. Ermert, F. D. Kelemen, P. Ludwig, and J. G. Pinto, 2017: Synoptic Analysis and Hindcast of an Intense Bow Echo in Western Europe: The 9 June 2014 Storm. *Weather and Forecasting*, 32(3), 1121-1141, <https://doi.org/10.1175/Waf-D-16-0192.1>
- McMurdie, L. A., A. K. Rowe, R. A. Houze, S. R. Brodzik, J. P. Zagrodnik, and T. M. Schuldt, 2018: Terrain-enhanced precipitation processes above the melting layer: Results from OLYMPEX. *Journal of Geophysical Research: Atmospheres*, 123(21), 12,194-112,209, <https://doi.org/10.1029/2018JD029161>



- Miao, J.-E., and M.-J. Yang, 2020: A modeling study of the severe afternoon thunderstorm event at Taipei on 14 June 2015: The roles of sea breeze, microphysics, and terrain. *J. Meteor. Soc. Japan*, 98, 129–152, <https://doi.org/10.2151/jmsj.2020-008>
- Mulholland, J. P., J. M. Peters, and H. Morrison, 2021: How does vertical wind shear influence entrainment in squall lines? *Journal of the Atmospheric Sciences*, 78(6), 1931–1946, <https://doi.org/10.1175/JAS-D-20-0299.1>
- Orlanski, I. 1975: A rational subdivision of scales for atmospheric processes. *Bulletin of the American Meteorological Society*, 527–530, <http://www.jstor.org/stable/26216020>
- Ooyama, K. V., 2002: The cubic-spline transform method: Basic definitions and tests in a 1D single domain. *Mon. Wea. Rev.*, 130, 2392–2415.
- Pan, Y.-T., and M.-J. Yang, 2022: Asymmetric Structures of a Squall-Line MCS over Taiwan with Significant Hydraulic Jumps. *Asia-Pacific Journal of Atmospheric Sciences*, 58(3), 415–433. <https://doi.org/10.1007/s13143-021-00262-1>
- Parker, M. D. (2010). Relationship between system slope and updraft intensity in squall lines. *Monthly Weather Review*, 138(9), 3572–3578, <https://doi.org/10.1175/2010MWR3441.1>
- Rotunno, R., J. B. Klemp, and M. L. Weisman, 1988: A Theory for Strong, Long-Lived Squall Lines. *Journal of the Atmospheric Sciences*, 45(3), 463–485, [https://doi.org/10.1175/1520-0469\(1988\)045<0463:Atfslw>2.0.Co;2](https://doi.org/10.1175/1520-0469(1988)045<0463:Atfslw>2.0.Co;2)
- Smull, B. F., and R. A. Houze, 1987: Rear Inflow in Squall Lines with Trailing Stratiform Precipitation. *Monthly Weather Review*, 115(12), 2869–2889, [https://doi.org/10.1175/1520-0493\(1987\)115<2869:Riislw>2.0.Co;2](https://doi.org/10.1175/1520-0493(1987)115<2869:Riislw>2.0.Co;2)
- Spuler, S. M., M. Hayman, R. A. Stillwell, J. Carnes, T. Bernatsky, and K. S. Repasky, 2021: MicroPulse DIAL (MPD)—a diode-laser-based lidar architecture for quantitative atmospheric profiling. *Atmospheric Measurement Techniques*, 14(6), 4593–4616, <https://doi.org/10.5194/amt-14-4593-2021>.

van Lier-Walqui, M., and Coauthors, 2016: On polarimetric radar signatures of deep convection for model evaluation: Columns of specific differential phase observed during MC3E. *Monthly Weather Review*, 144(2), 737-758, <https://doi.org/10.1175/MWR-D-15-0100.1>.



Wakimoto, R. M., P. Stauffer, and W.-C. Lee, 2015: The Vertical Vorticity Structure within a Squall Line Observed during BAMEX: Banded Vorticity Features and the Evolution of a Bowing Segment. *Monthly Weather Review*, 143(1), 341-362, <https://doi.org/10.1175/Mwr-D-14-00246.1>

Weisman, M. L. 1992: The Role of Convectively Generated Rear-Inflow Jets in the Evolution of Long-Lived Mesoconvective Systems. *Journal of the Atmospheric Sciences*, 49(19), 1826-1847, [https://doi.org/10.1175/1520-0469\(1992\)049<1826:Trocgr>2.0.Co;2](https://doi.org/10.1175/1520-0469(1992)049<1826:Trocgr>2.0.Co;2)

Weisman, M. L. 1993: The Genesis of Severe, Long-Lived Bow Echoes. *Journal of the Atmospheric Sciences*, 50(4), 645-670, [https://doi.org/10.1175/1520-0469\(1993\)050<0645:Tgosll>2.0.Co;2](https://doi.org/10.1175/1520-0469(1993)050<0645:Tgosll>2.0.Co;2)

Weisman, M. L. 2001: Bow echoes: A tribute to TT Fujita. *Bulletin of the American Meteorological Society*, 82(1), 97-116, [https://doi.org/10.1175/1520-0477\(2001\)082<0097:BEATT>2.3.CO;2](https://doi.org/10.1175/1520-0477(2001)082<0097:BEATT>2.3.CO;2)

Weisman, M. L., and C. A. Davis, 1998: Mechanisms for the generation of mesoscale vortices within quasi-linear convective systems. *Journal of the Atmospheric Sciences*, 55(16), 2603-2622, [https://doi.org/10.1175/1520-0469\(1998\)055<2603:Mftgom>2.0.Co;2](https://doi.org/10.1175/1520-0469(1998)055<2603:Mftgom>2.0.Co;2)

Weisman, M. L., and R. Rotunno, 2004: "A theory for strong long-lived squall lines" revisited. *Journal of the Atmospheric Sciences*, 61(4), 361-382, [https://doi.org/10.1175/1520-0469\(2004\)061<0361:ATFSLS>2.0.CO;2](https://doi.org/10.1175/1520-0469(2004)061<0361:ATFSLS>2.0.CO;2)

Weisman, M. L., and R. J. Trapp, 2003: Low-level mesovortices within squall lines and bow echoes. Part I: Overview and dependence on environmental shear. *Monthly Weather Review*, 131(11), 2779-2803, [https://doi.org/10.1175/1520-0493\(2003\)131<2779:Lmwsla>2.0.Co;2](https://doi.org/10.1175/1520-0493(2003)131<2779:Lmwsla>2.0.Co;2)

Xie, J., and R. A. Scofield, 1989: Satellite-derived rainfall estimates and propagation characteristics associated with mesoscale convective systems (MCSs).
<https://repository.library.noaa.gov/view/noaa/19349>

Yang, M. J., and R. A. Houze, 1995: Multicell Squall-Line Structure as a Manifestation of Vertically Trapped Gravity-Waves. *Monthly Weather Review*, 123(3), 641-661,
[https://doi.org/10.1175/1520-0493\(1995\)123<0641:Mslsaa>2.0.Co;2](https://doi.org/10.1175/1520-0493(1995)123<0641:Mslsaa>2.0.Co;2)

Yuter, S. E., and R. A. Houze, 1995: Three-dimensional kinematic and microphysical evolution of Florida cumulonimbus. Part II: Frequency distributions of vertical velocity, reflectivity, and differential reflectivity. *Monthly Weather Review*, 123(7), 1941-1963,
[https://doi.org/10.1175/1520-0493\(1995\)123<1941:TDKAME>2.0.CO;2](https://doi.org/10.1175/1520-0493(1995)123<1941:TDKAME>2.0.CO;2)

Zhou, A., K. Zhao, W.-C. Lee, H. Huang, D. Hu, and P. Fu, 2020: VDRAS and Polarimetric Radar Investigation of a Bow Echo Formation After a Squall Line Merged With a Preline Convective Cell. *Journal of Geophysical Research-Atmospheres*, 125(7), <https://doi.org/10.1029/2019JD031719>

Tables

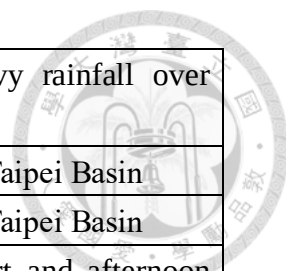


Table 1. Mean convective available potential energy (CAPE) and environmental vertical wind shear (VWS) in the lower and middle troposphere calculated from the soundings launched in Hsinchu during TAHOPE.

Time	CAPE	10m-2km VWS	10m-3km VWS	10m-5km VWS
00 UTC	3233.8 J/kg	3.5m/s	2.5m/s	2.5m/s
06 UTC	3937.7 J/kg	5.9m/s	4.3m/s	7.4m/s

Table 2. TAHOPE 2022 Experiment Period IOP and SOP List. (adapted from 楊 等, 2023)

Number	Period	Description
SOP1	5/25 00Z – 5/26 00Z	Meiyu front and mesoscale convective systems
IOP1	5/26 00Z – 5/28 00Z	Meiyu front and mesoscale convective systems (Back-building convective system at 5/26)
IOP2	5/31 00Z – 5/31 12Z	Afternoon thunderstorm (Observation in Yilan)
IOP3	6/06 06Z – 6/12 12Z	Meiyu front and MCS (Back-building in Miaoli; squall line approaching Taiwan)
IOP4	6/15 00Z – 6/16 00Z	Weakened southwesterly flow ahead of the Meiyu front (low-level dry air intrusion into northern Taiwan)
SOP2	6/23 00Z – 6/24 00Z	Afternoon thunderstorm
SOP3	6/24 00Z – 6/25 00Z	Afternoon thunderstorm (hail observed in the Taipei Basin)
IOP5	6/25 00Z – 6/25 15Z	Afternoon thunderstorm (intense convective cells observed in Taipei, Taoyuan, and Miaoli)
SOP4	6/29 00Z – 6/30 00Z	Afternoon thunderstorm
IOP6	7/01 00Z – 7/04 00Z	Rainfall associated with Typhoons Chaba and Aere (heavy rainfall in eastern and southern Taiwan)
IOP7	7/05 00Z – 7/05 04Z	Typhoon Aere approached Kyushu, Japan (heavy rainfall in Kyushu and Honshu, Japan)
SOP5	7/05 00Z – 7/06 00Z	Afternoon thunderstorm in the Taipei Basin
SOP6	7/06 00Z – 7/07 00Z	Afternoon thunderstorm in the Taipei Basin



IOP8	7/13 06Z – 7/16 06Z	TD 90W brings rainfall (heavy rainfall over eastern and southern Taiwan)
SOP7	7/19 00Z – 7/20 00Z	Afternoon thunderstorm in the Taipei Basin
SOP8	7/20 00Z – 7/21 00Z	Afternoon thunderstorm in the Taipei Basin
IOP9	7/29 00Z – 7/29 12Z	Tropical low moisture transport and afternoon convection (heavy rain in the evening; thunderstorms and lightning in northern Taiwan)
IOP10	8/01 00Z – 8/03 18Z	Tropical low moisture transport (heavy rainfall occurring in the Taipei Basin, central Taiwan, and southern Taiwan)
IOP11	8/04 00Z – 8/04 12Z	Afternoon thunderstorm in the Taipei Basin

Figures

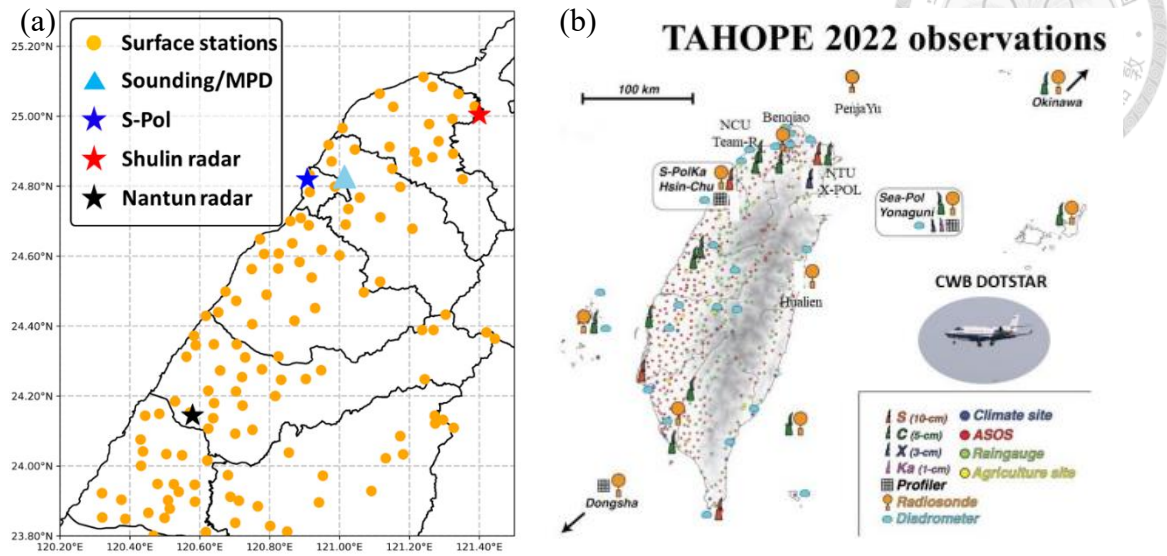


Figure 1. (a) Distribution of surface observation instruments and radar sites in the study area during TAHOPE. (b) The configuration of observational instruments used in the field experiment is illustrated in Fig. 1, adapted from 楊明仁等 (2023).

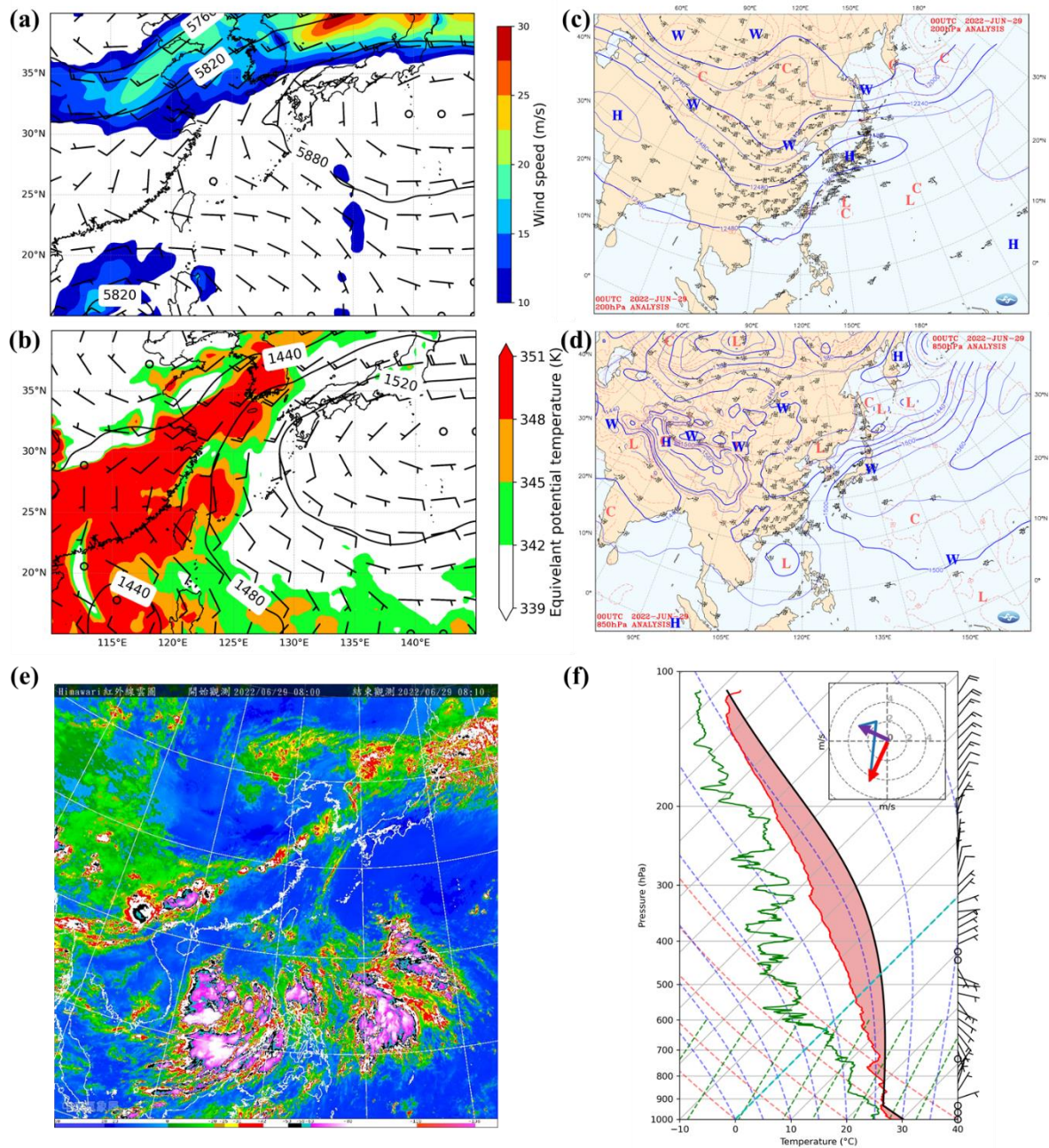


Figure 2. (a) 500-hPa geopotential height, wind vectors, and wind speed (shaded) at 08 LST on 29 June 2022, derived from the ERA5 reanalysis. (b) The corresponding 850-hPa geopotential height, wind vectors, and equivalent potential temperature (θ_e). (c) 850-hPa synoptic analysis chart. (d) 200-hPa synoptic analysis chart. (e) HIMAWARI IR color image for Asia area. (f) 00 UTC sounding launched by the CSU in Hsinchu. The shaded area represents the CAPE. The wind rose in the top-right corner displays vertical wind shear vectors: red and purple arrows indicate 10 m–2 km and 10 m–5 km shear, respectively.

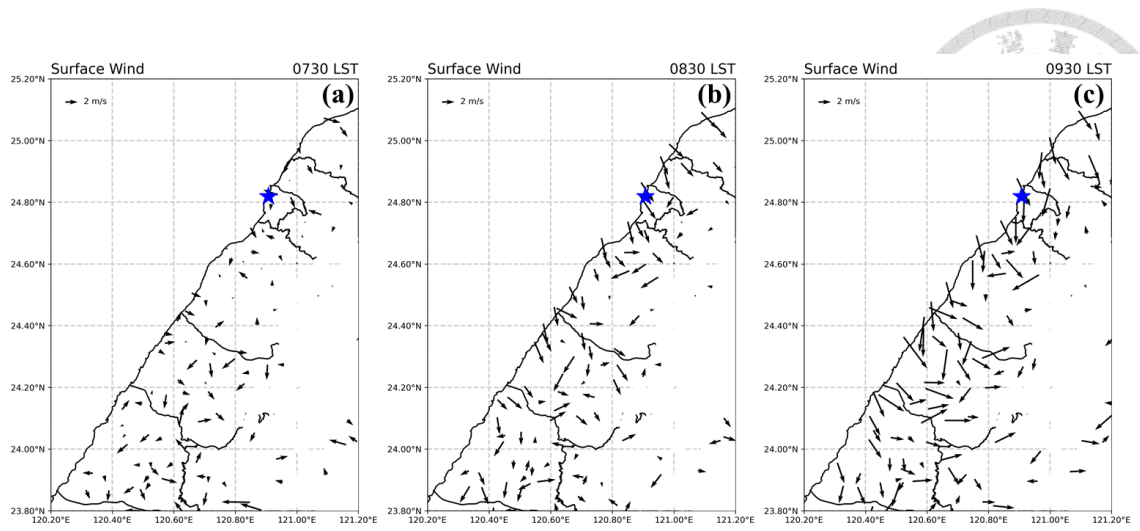


Figure 3. Surface wind observations from ground-based stations at (a) 0730 LST, (b) 0830 LST, and (c) 0930 LST.

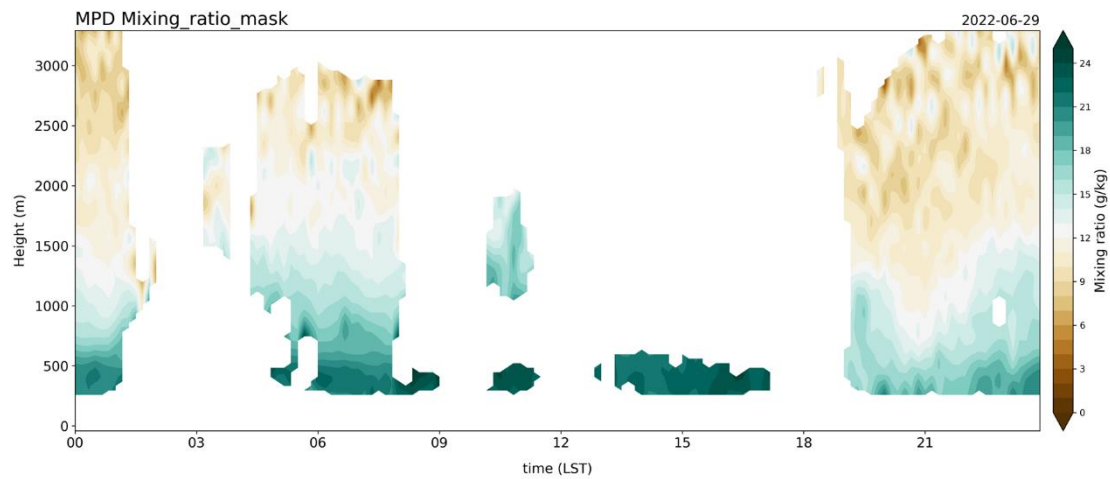


Figure 4. Low-level water vapor content observed in Hsinchu by the MicroPulse Differential Absorption Lidar (MPD) and converted to vapor mixing ratio (units: g/kg) using the atmospheric density derived from Hsinchu radiosonde observations.

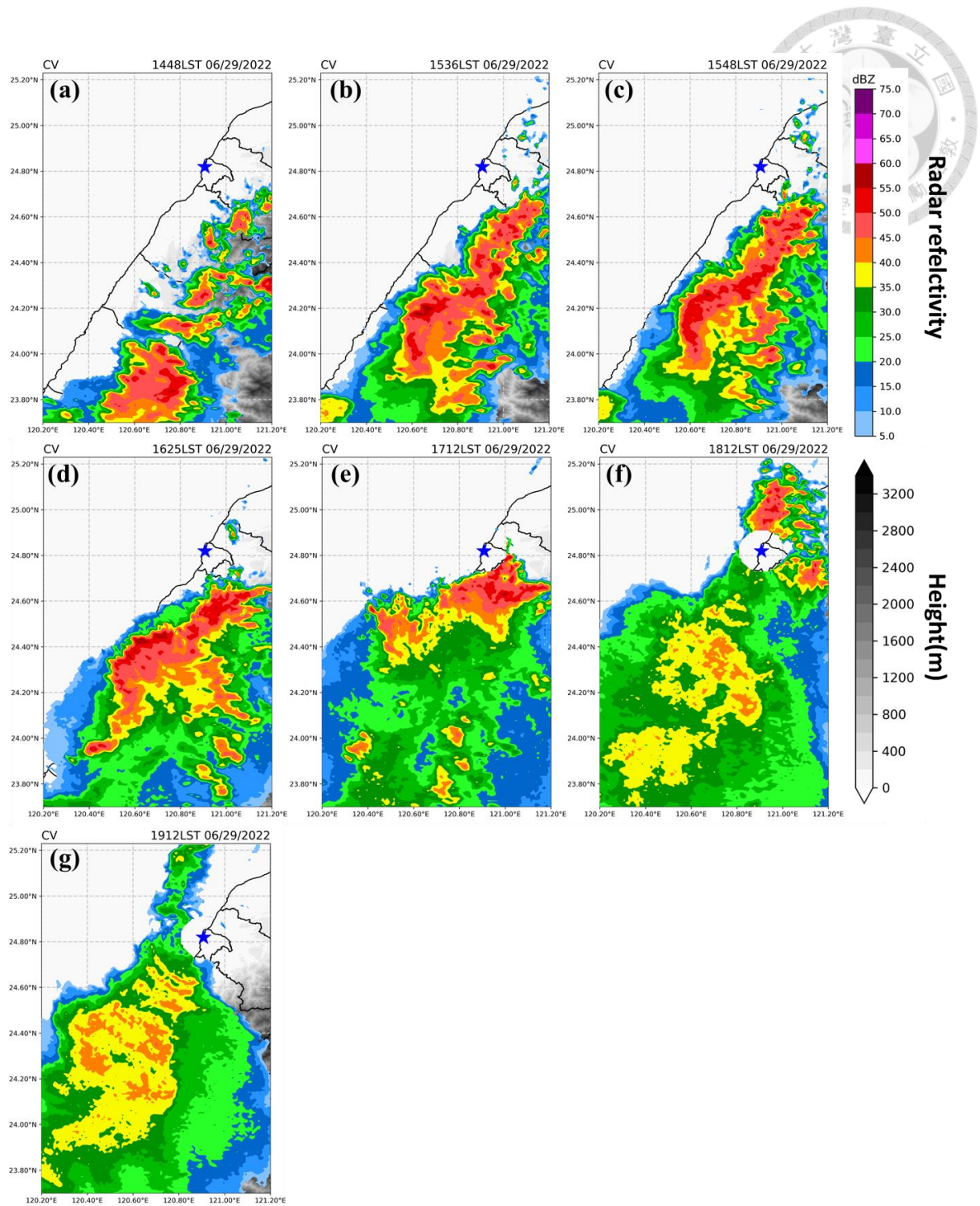


Figure 5. Composite radar reflectivity from the S-Pol at 2-km altitude from 1448 to 1912 LST. The blue star indicates the location of the S-Pol radar site. Panels: (a) 1448, (b) 1536, (c) 1548, (d) 1625, (e) 1712, (f) 1812, and (g) 1912 LST.

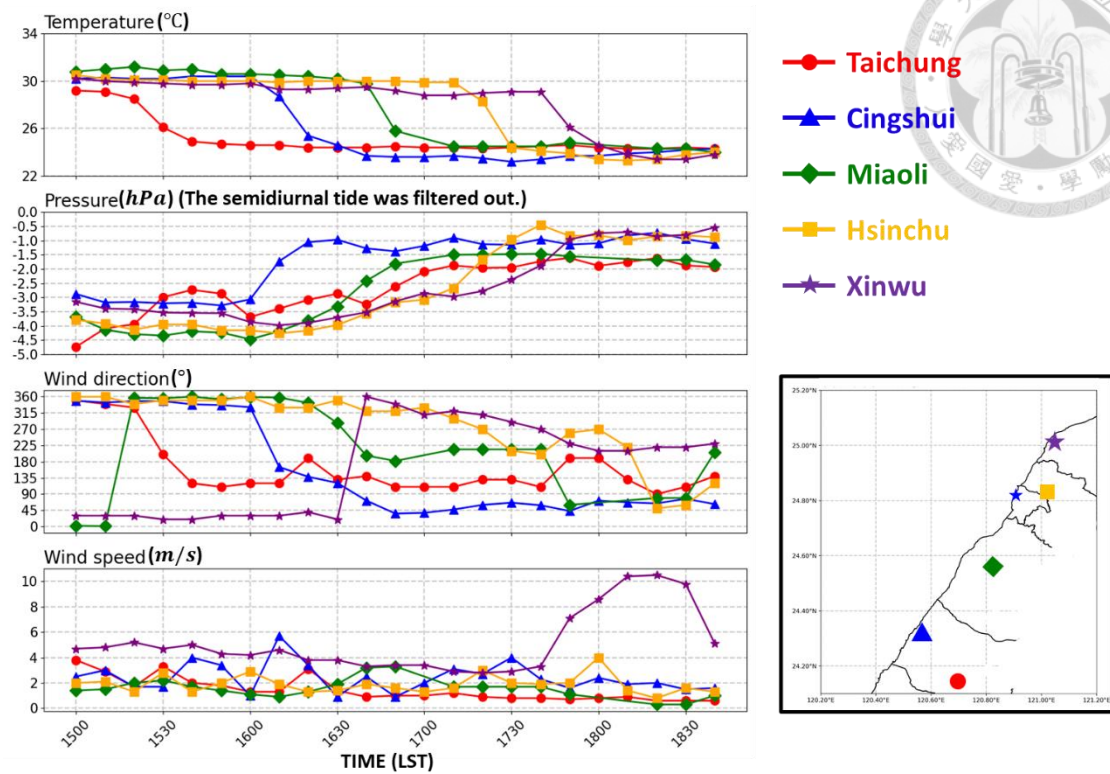


Figure 6. Major surface observation stations along the convective system's south-to-north propagation path, including Taichung, Qingshui, Miaoli, Hsinchu, and Xinwu. The panel at the bottom right shows the geographic locations of these stations. The time series plots on the left display, from top to bottom, are temperature, pressure (with the semi-diurnal tide removed), wind direction, and wind speed respectively.

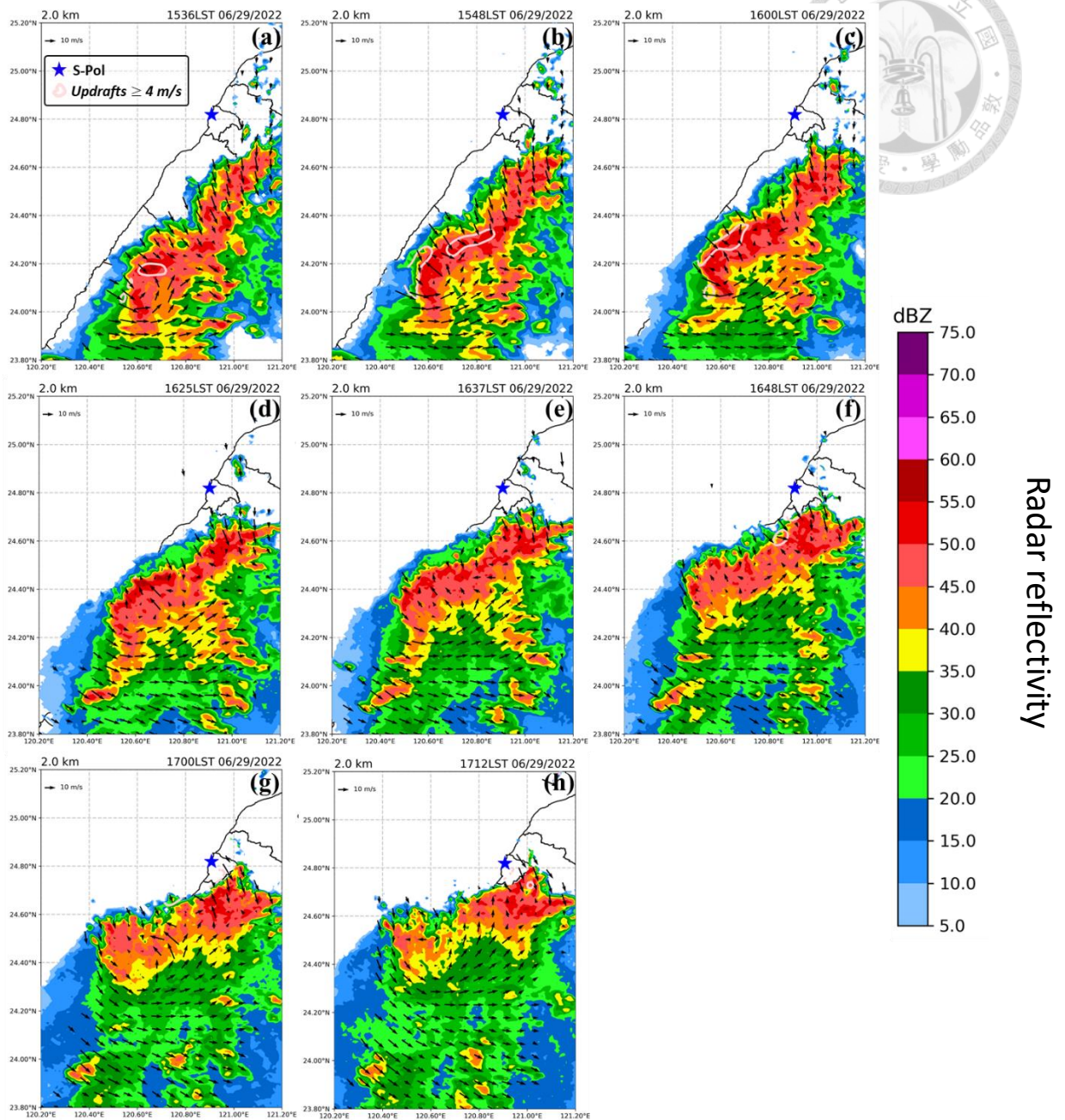


Figure 7. Composite reflectivity in units of dBZ at 2 km altitude and horizontal storm-relative wind (SRWIND) from 1536 to 1712 LST. The blue star indicates the location of the S-Pol radar. Pink contours mark regions where vertical velocity exceeds 4 m s^{-1} at 2-km height, highlighting areas of strong low-level updrafts. Panel: (a) 1536, (b) 1548, (c) 1600, (d) 1625, (e) 1637, (f) 1648, (g) 1700, (h) 1712 LST

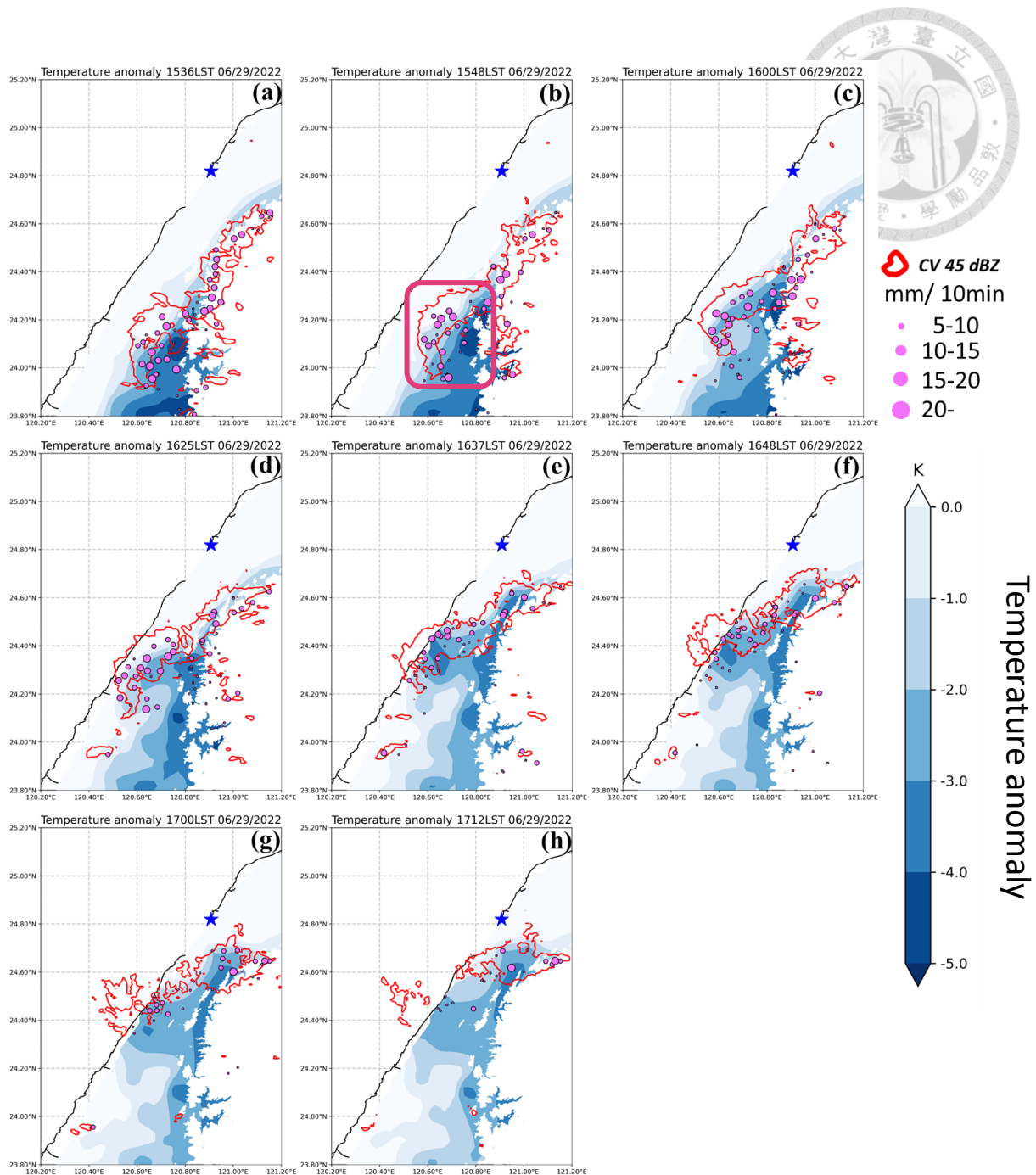


Figure 8. Temperature anomaly (in units of K) overlaid with 45 dBZ radar reflectivity contours and 10-minute accumulated rainfall from 1536 to 1712 LST. The size of the pink circles is proportional to the magnitude of accumulated rainfall; the largest circle represents a maximum accumulation of 20 mm per 10 minutes. Panel: (a) 1536, (b) 1548, (c) 1600, (d) 1625, (e) 1637, (f) 1648, (g) 1700, (h) 1712 LST

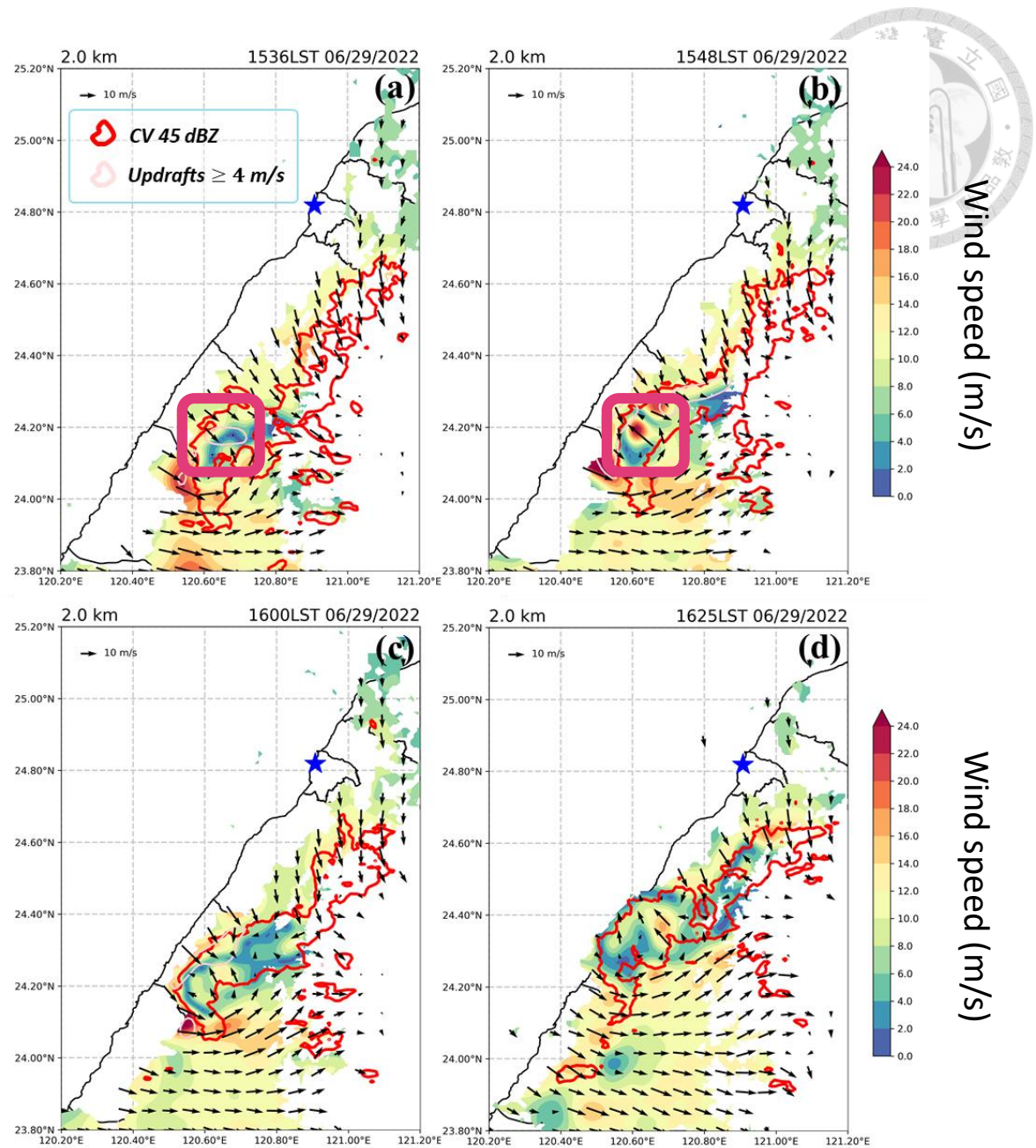


Figure 9. Storm-relative wind speed (units: $m s^{-1}$) and wind vectors overlaid with 45 dBZ radar reflectivity contours. Pink contours indicate regions where wind speed increases. The blue star marks the location of the S-Pol radar. Pink contours also denote areas where vertical velocity exceeds $4 m s^{-1}$ at 2-km height. Panel: (a) 1536, (b) 1548, (c) 1600, (d) 1625.

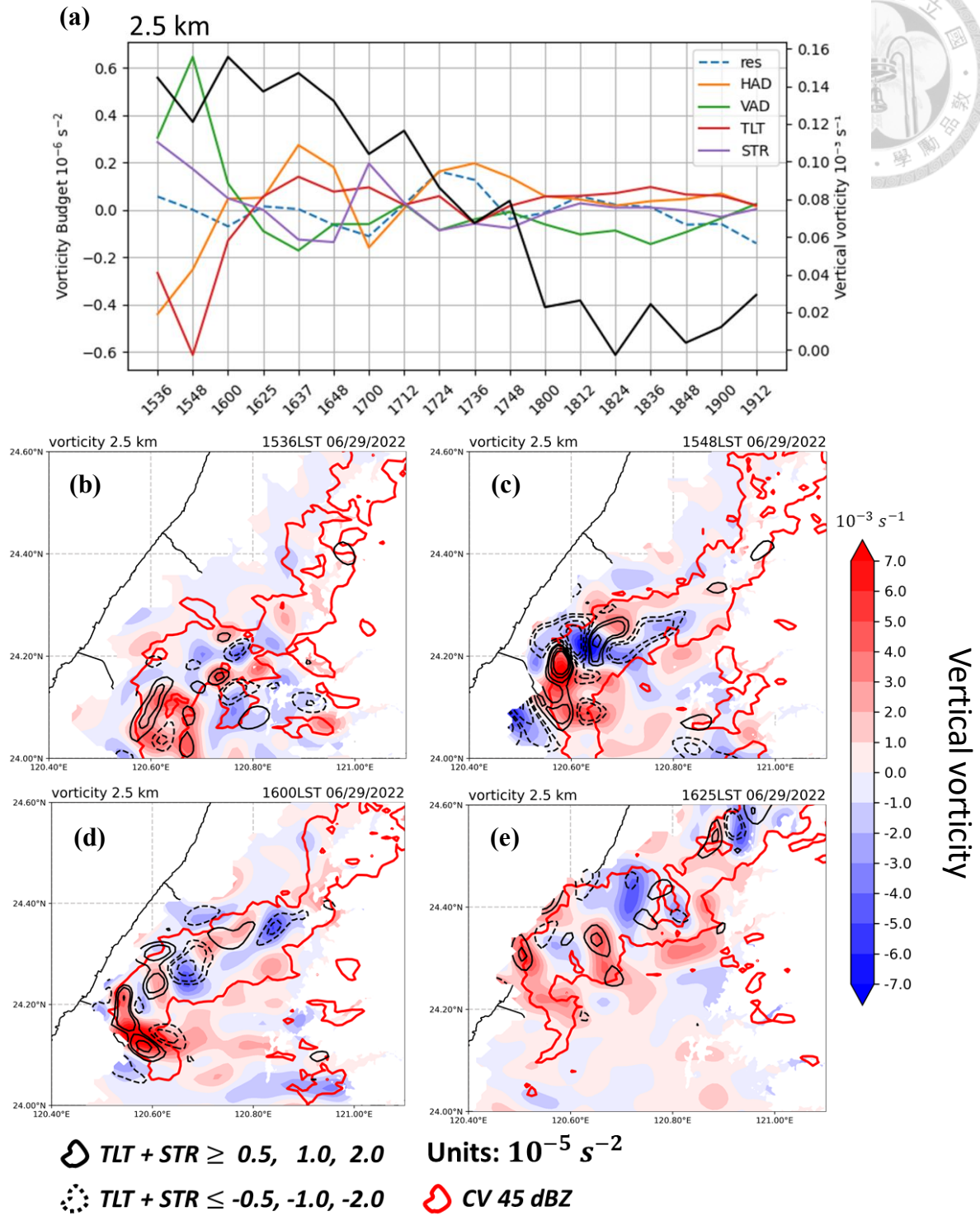


Figure 10. (a) Domain-averaged vorticity budget terms, with different colored lines representing individual contributions from each term. (b)-(e) vorticity (shaded; units: 10^{-3} s^{-1}) and the sum of the stretching and tilting terms from the vertical vorticity tendency equation (contours; units: 10^{-5} s^{-2}). Solid contours indicate positive tendencies (0.5, 1.0, and 2.0), while dashed contours represent negative tendencies (-0.5 , -1.0 , and -2.0). Panels (b)–(e) correspond to 1536, 1548, 1600 and 1625 LST, respectively.

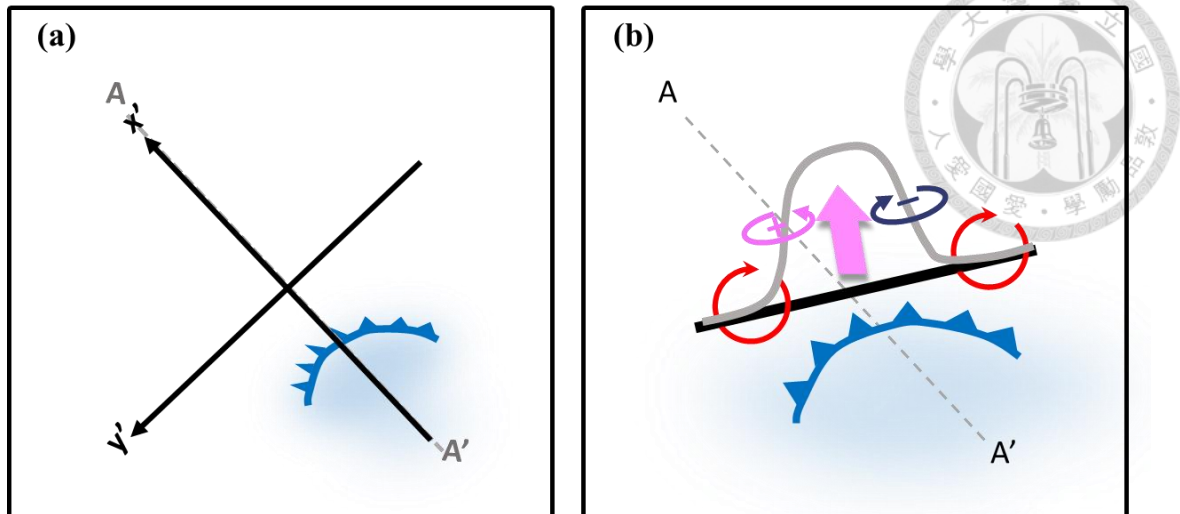


Figure 11. (a) Schematic illustration of the coordinate system rotation. (b) Conceptual diagram showing the tilting and stretching of horizontal vorticity.

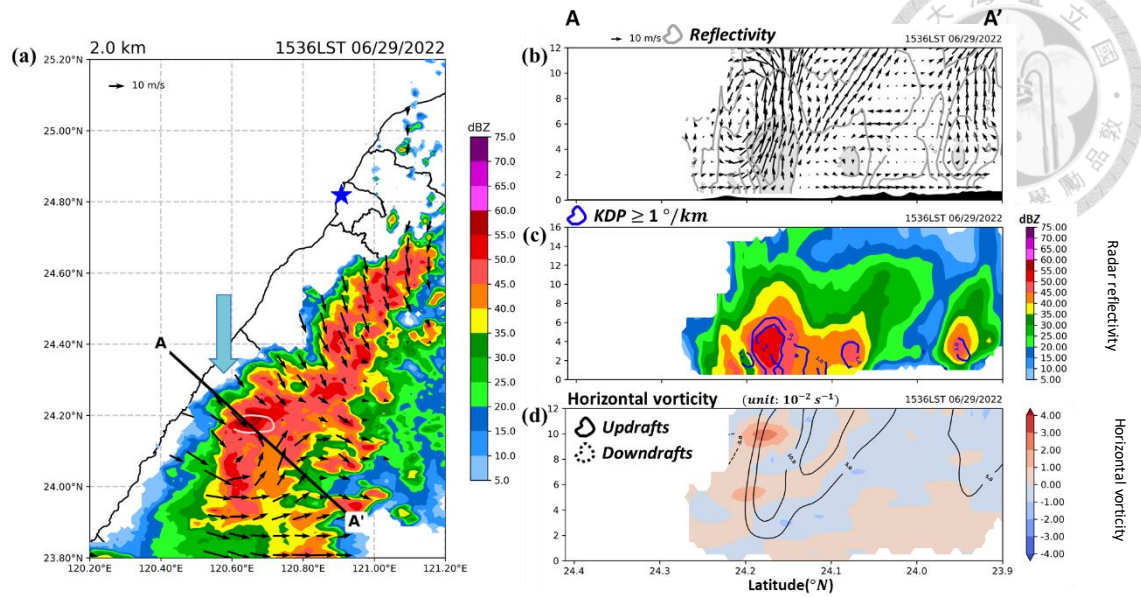


Figure 12. Cross-sectional analysis of the convective system at 1536 LST. (a) Location of cross section AA' overlaid on the composite radar reflectivity, indicating the system's position at this time. (b) Wind field along the AA' cross section, with the vertical component exaggerated by a factor of two. Contours represent radar reflectivity, with shading applied for values exceeding 45 dBZ. (c) Shaded colors represent radar reflectivity; contours denote regions where $K_{DP} \geq 1^\circ \text{ km}^{-1}$. (d) Horizontal vorticity projected onto the rotated cross-sectional plane (following the coordinate system shown in Fig. 11a); contours indicate vertical velocity.

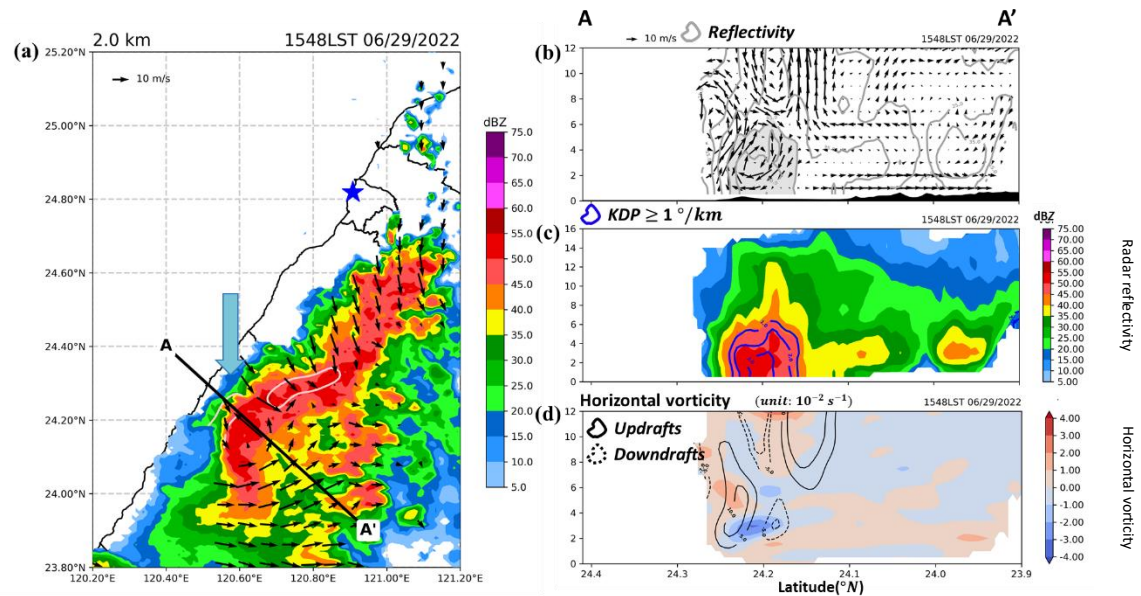


Figure 13. As in Fig. 12, but showing the cross-sectional analysis at 1548 LST.

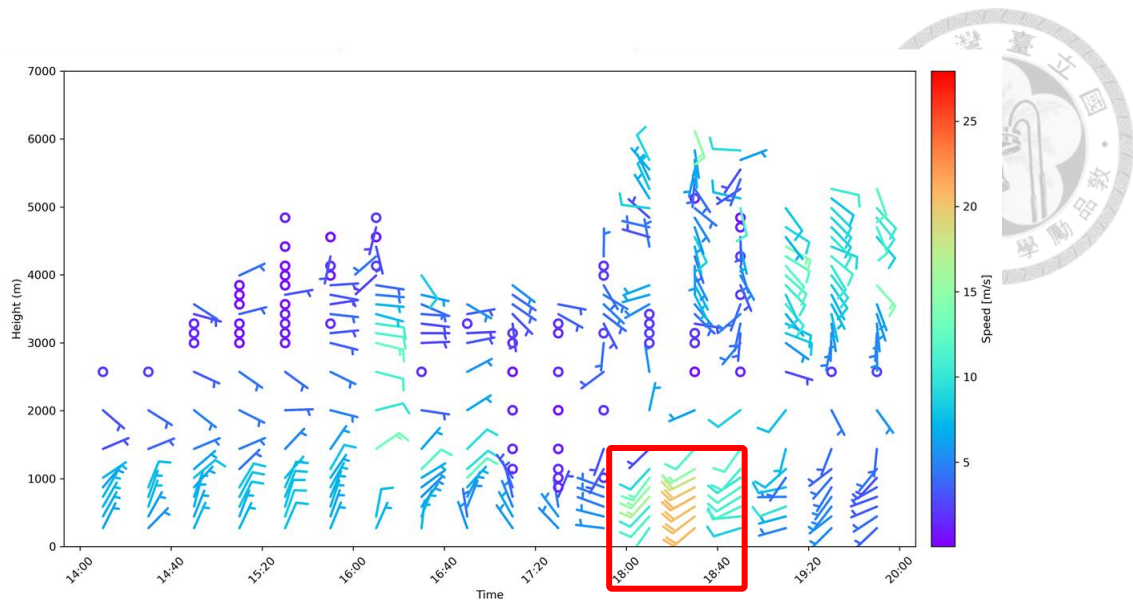


Figure 14. Cross-section of wind direction and speed derived from the Xinwu wind profiler between 1400 and 2000 LST. Wind barbs indicate wind direction and speed, and the background color shading represents wind speed magnitude as shown in colorbar. The red box marks the time period when changes in wind direction and speed begin to appear.

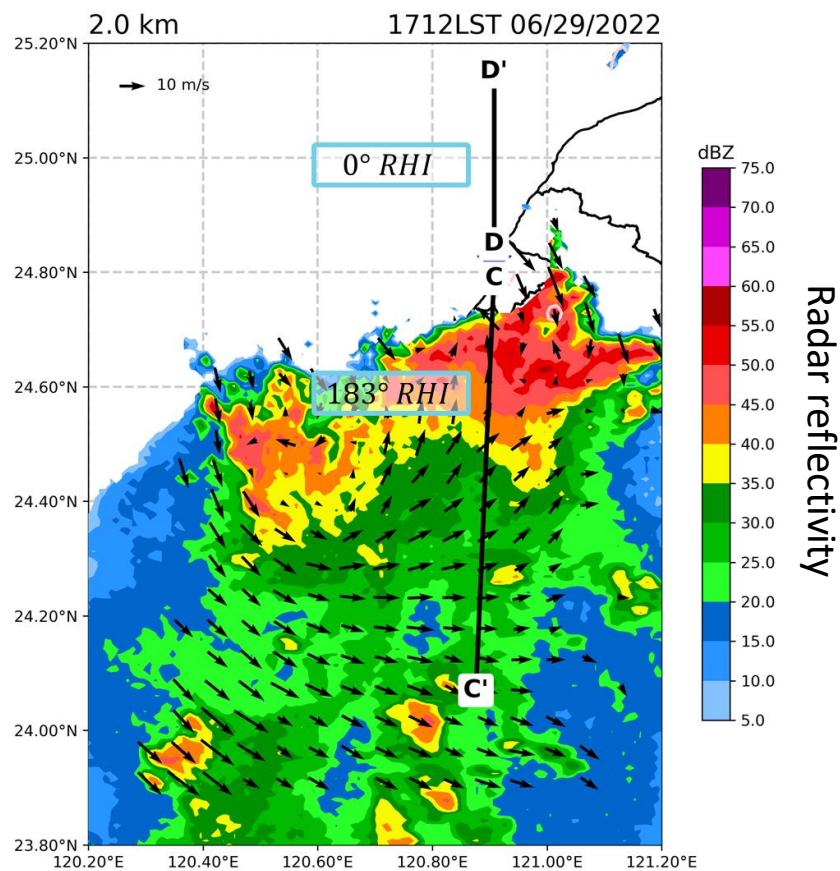


Figure 15. Locations of RHI cross-sections at azimuth angles of 183° and 0°.

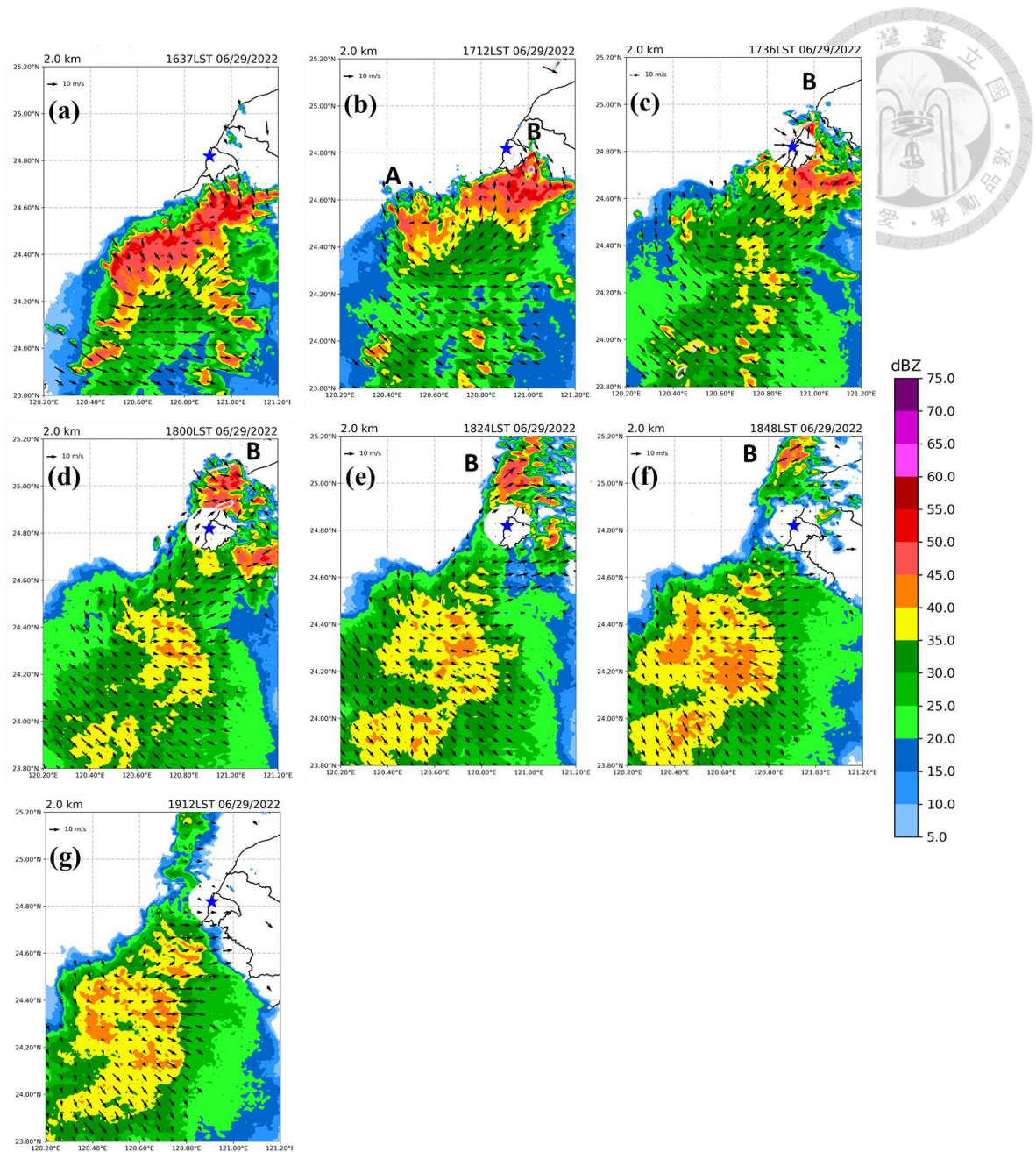


Figure 16. Same as Fig. 7, but for a different time period. Star indicates the location of S-Pol radar site. Panel: (a) 1637, (b) 1712, (c) 1736, (d) 1800, (e) 1824, (f) 1848, (g) 1912, LST.

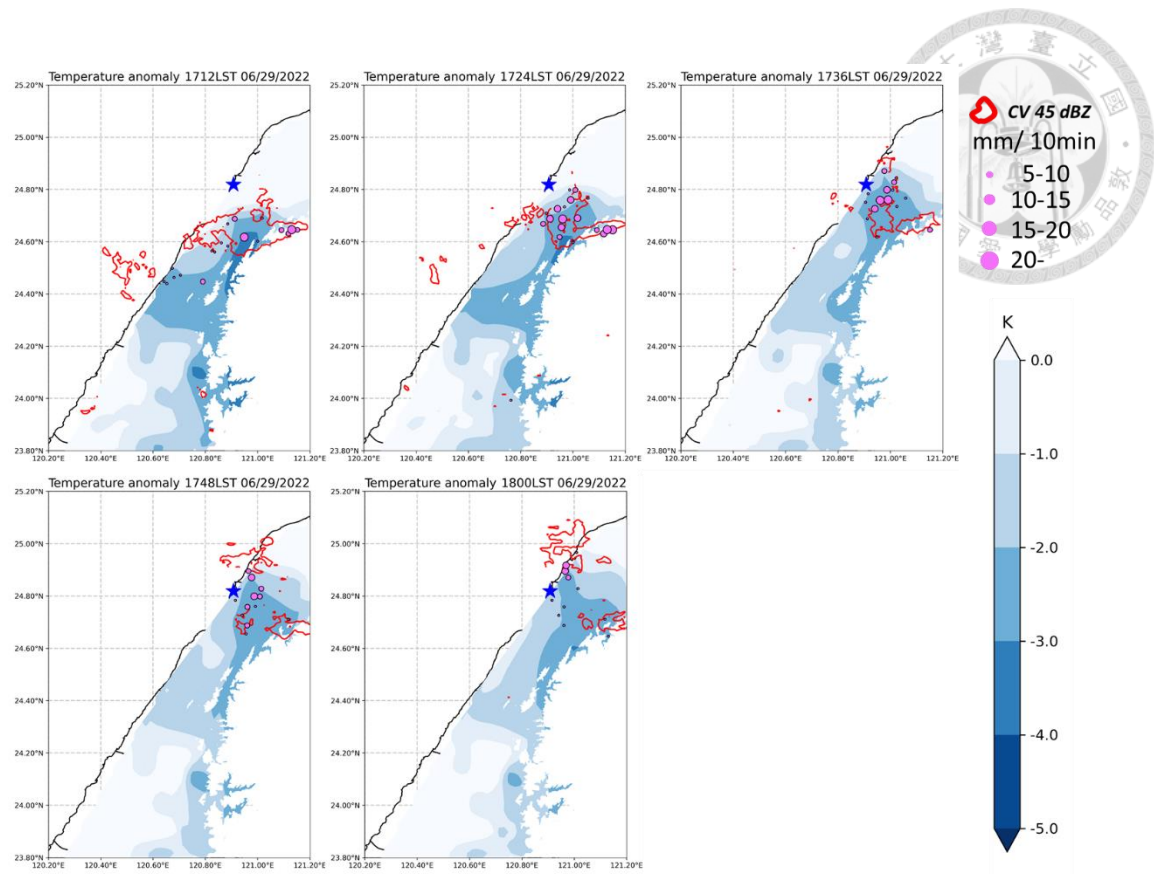


Figure 17. Same as Fig. 8, but for a different time period. Panel: (a) 1712, (b) 1724, (c) 1736, (d) 1748, (e) 1800 LST.

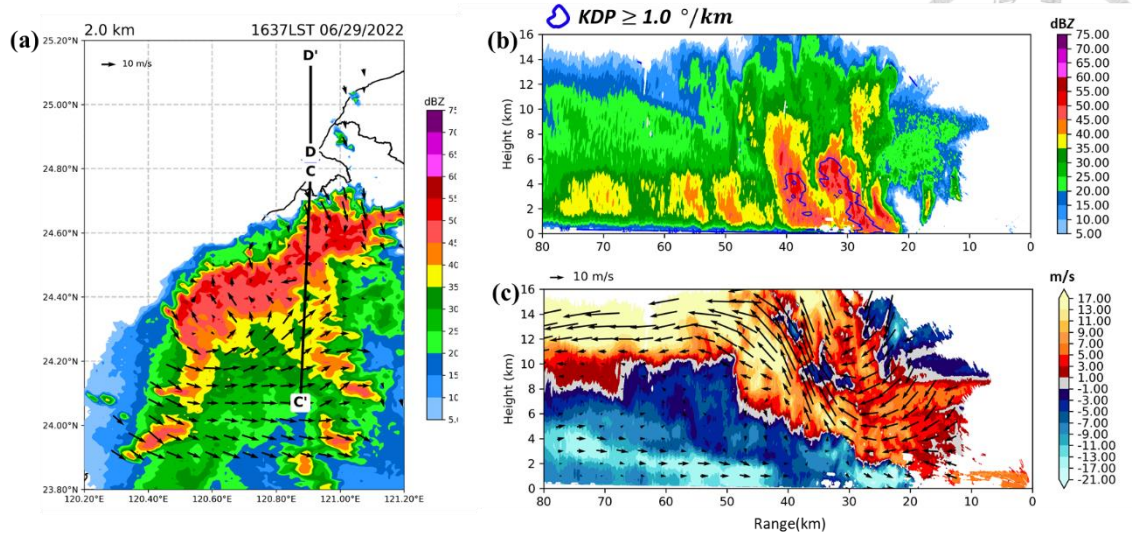


Figure 18. RHI radar analysis. (a) PPI image showing the location of the convective system and the orientation of the RHI cross section at 1637 LST. (b) RHI cross section of reflectivity with overlaid contours of $K_{DP} \geq 1.0$ ° km⁻¹. (c) RHI cross section of radial velocity and wind vectors along the plane of the cross-section, with vertical velocity exaggerated by a factor of 2.

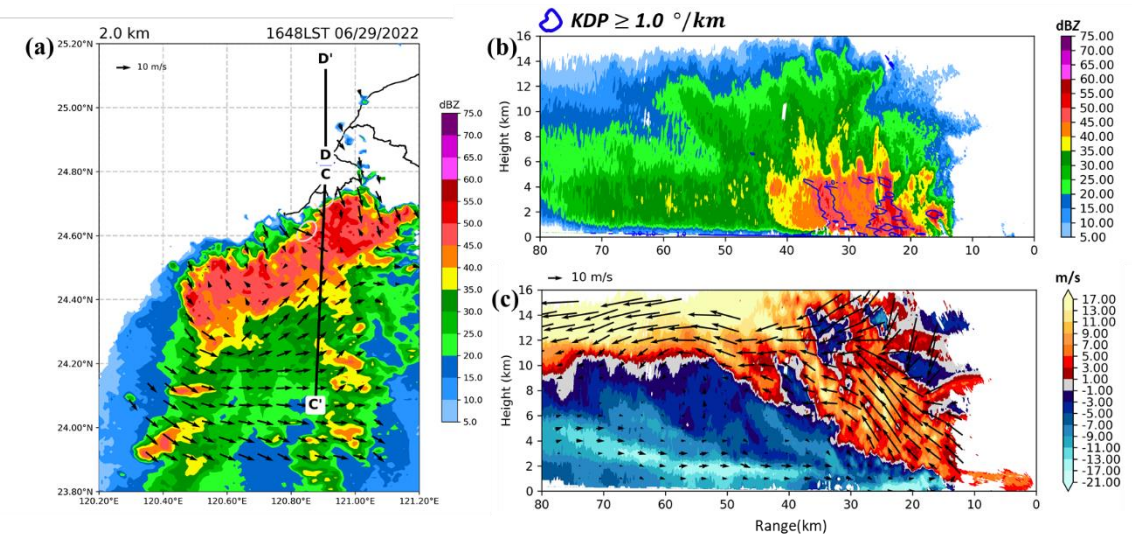


Figure 19. Same as Fig. 18, but at 1648 LST.

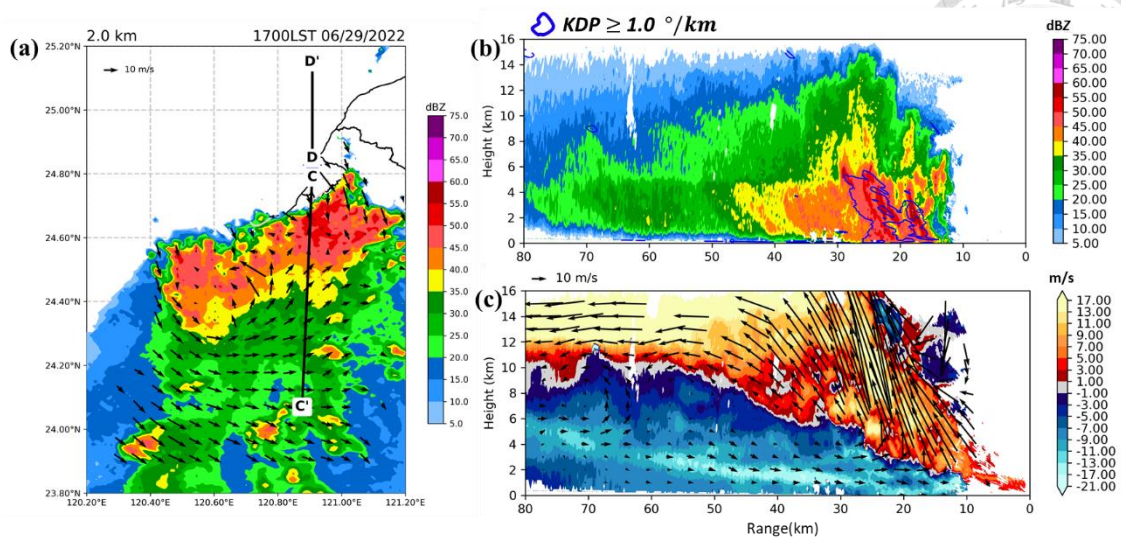


Figure 20. Same as Fig. 18, but at 1700 LST.

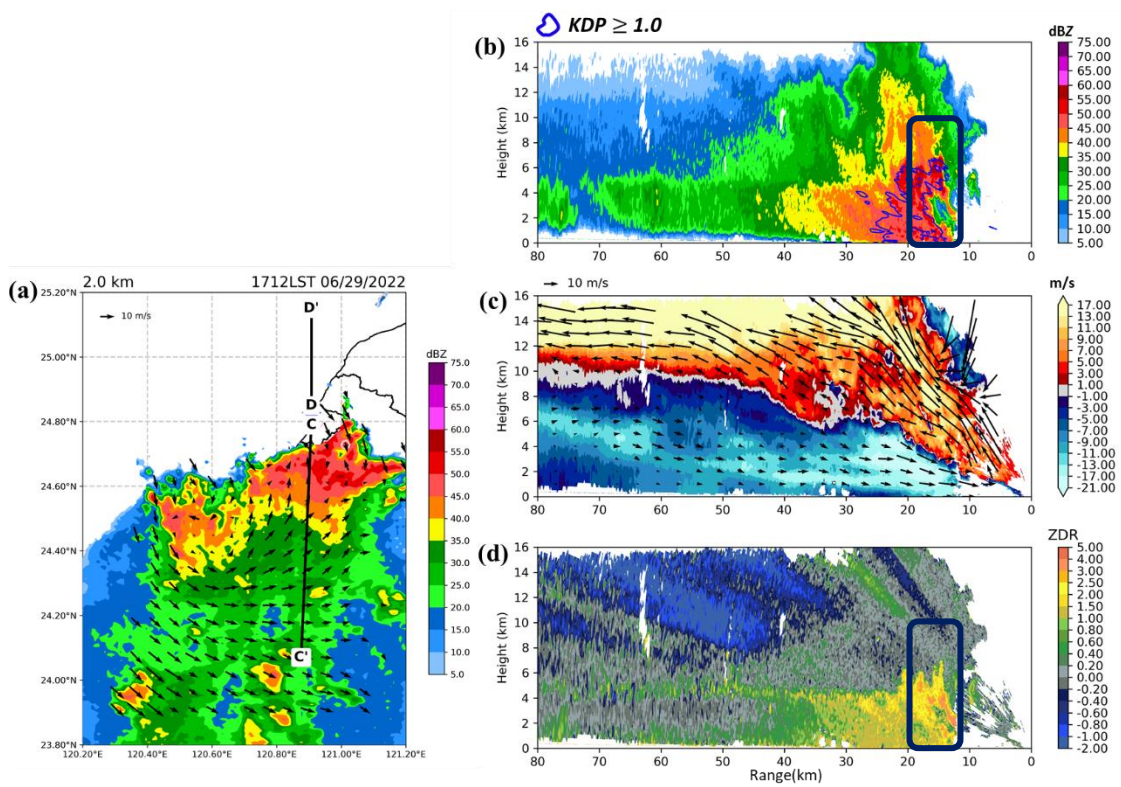


Figure 21. (a)-(c) Same as Fig. 18, but at 1712 LST. (d) RHI cross-section of ZDR. The dark indigo box highlights the region where the K_{DP} column extends above the melting layer (4–5 km) and where both DBZ and Z_{DR} signals are weak.

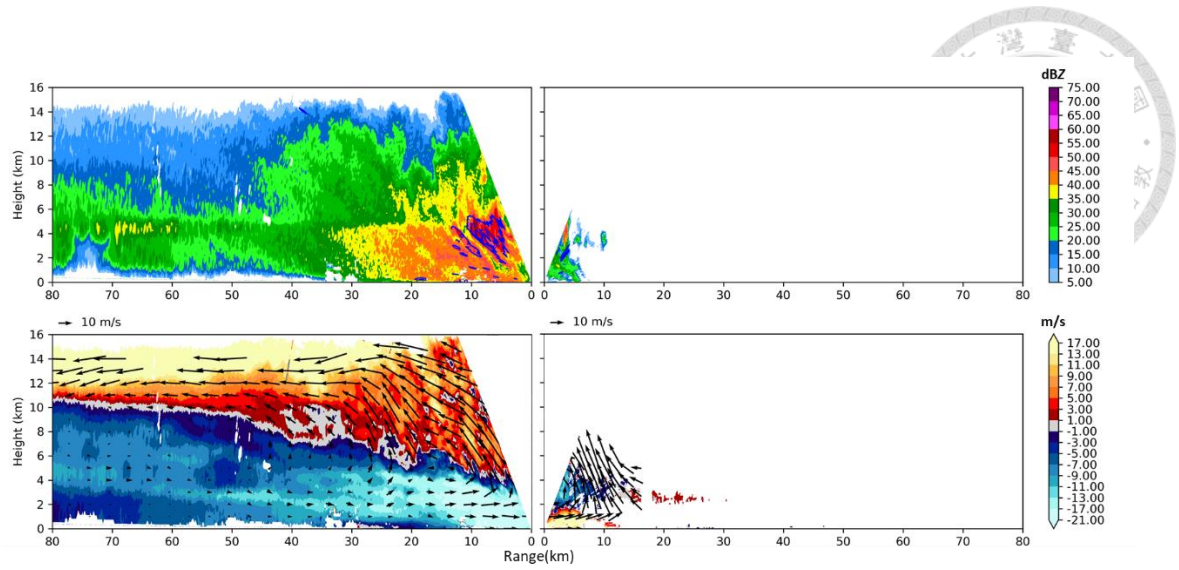


Figure 22. Left panels show the 183° RHI cross sections, and right panels show the 0° RHI cross sections. From top to bottom, the panels display radar reflectivity overlaid with areas where $K_{DP} \geq 1^\circ \text{ km}^{-1}$, and radial velocity overlaid with horizontal wind vectors. The time is at 1736 LST.

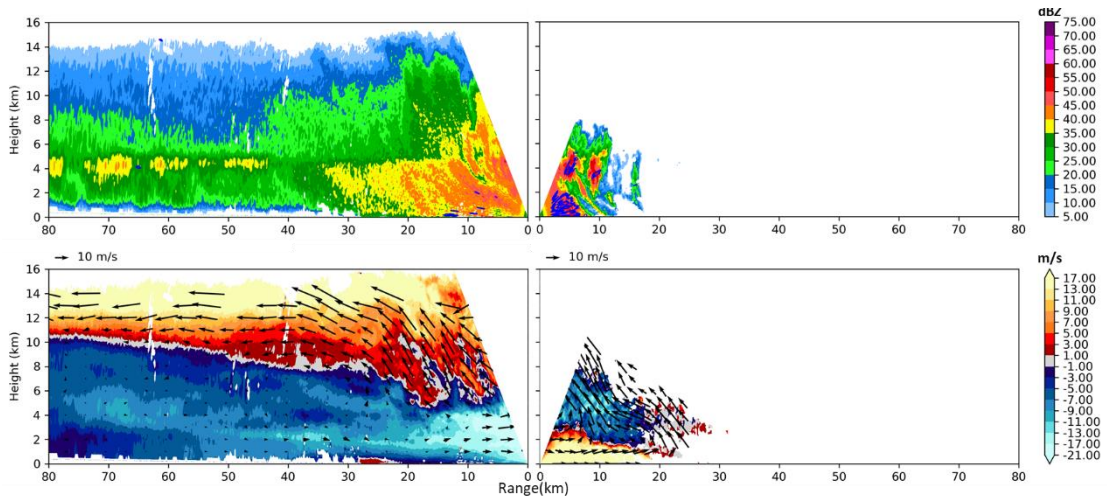


Figure 23. Same as Fig. 22, but at 1748 LST.

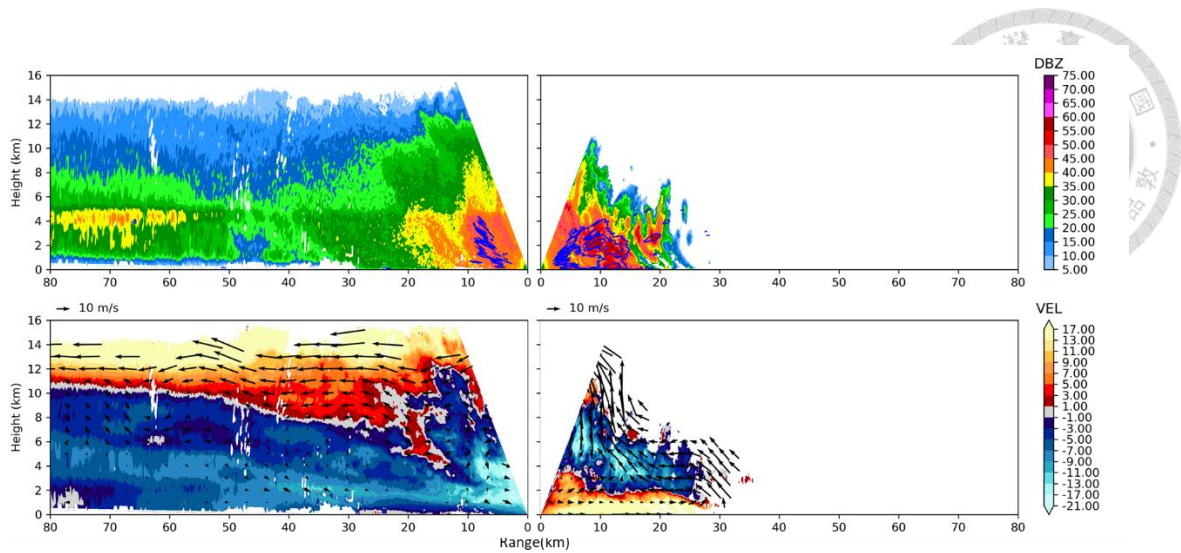


Figure 24. Same as Fig. 22, but at 1800 LST.

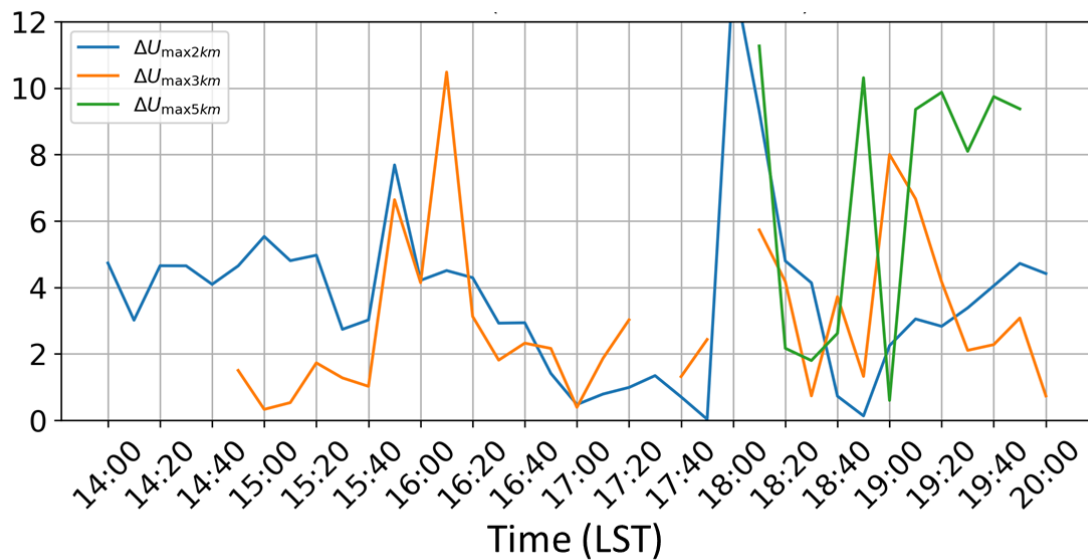


Figure 25. Vertical shear of horizontal wind in the lower to mid-troposphere, as derived from the Xinwu wind profiler. The blue line indicates vertical shear between 10 m and 2 km, the orange line represents shear up to 3 km, and the green line shows shear up to 5 km (note that some data are missing).

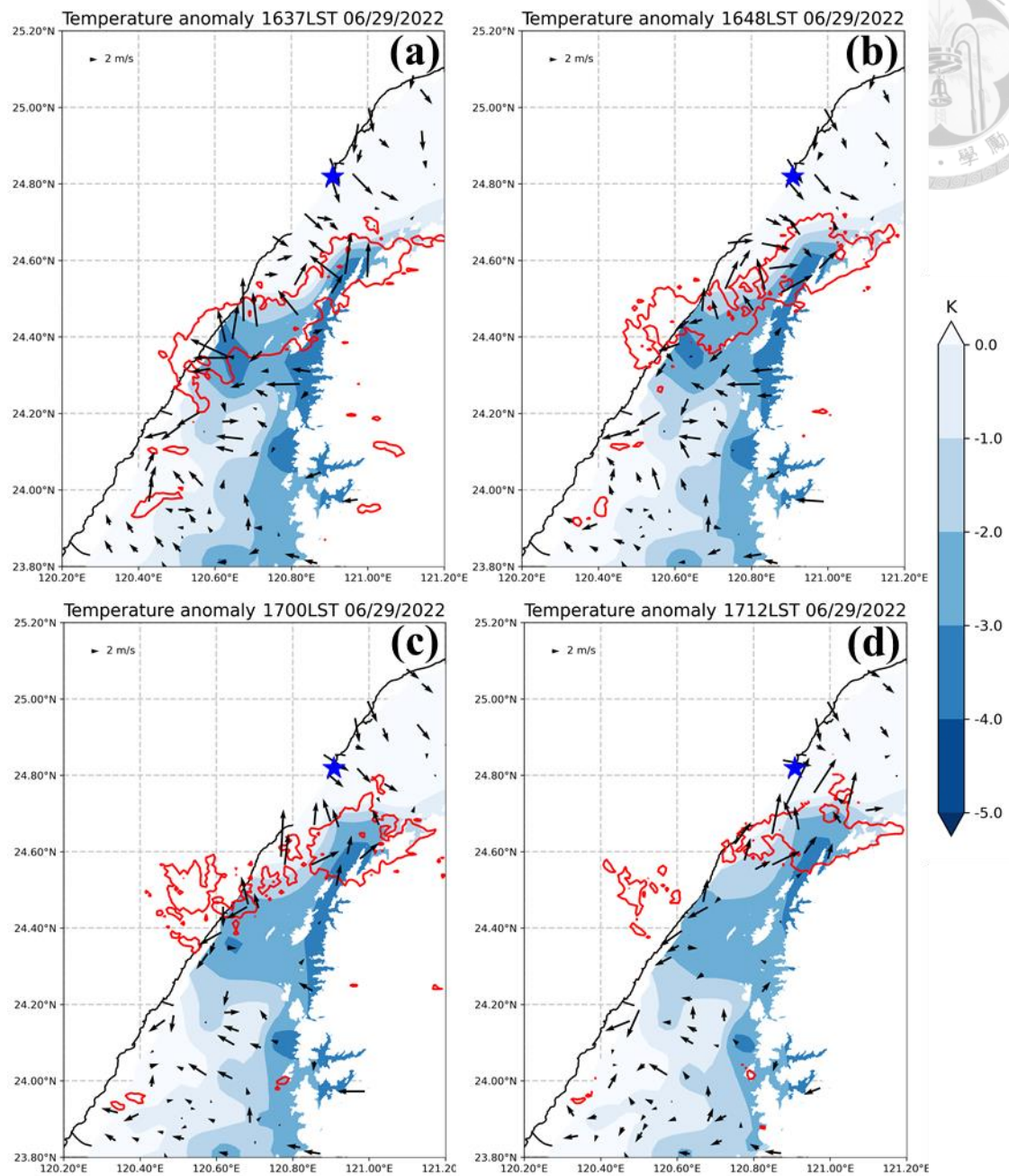


Figure 26. Temperature anomaly (in units of K) overlaid with 45 dBZ radar reflectivity contours and surface wind from (a)16:37, (b) 16:48, (c) 17:00 to (d)17:12 LST.

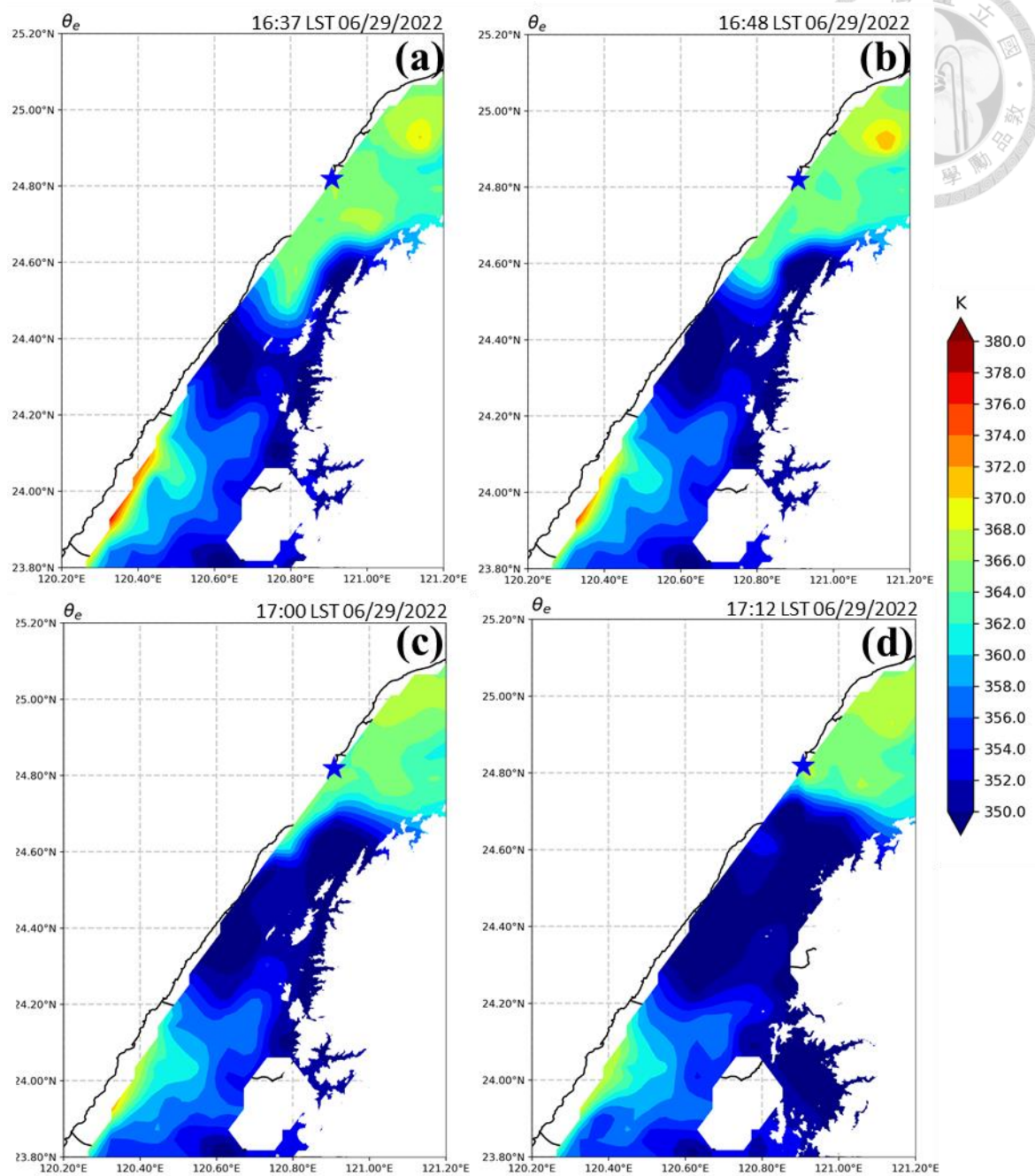


Figure 27. Surface equivalent potential (in units of K) temperature from (a)16:37, (b) 16:48, (c) 17:00 to (d)17:12 LST.

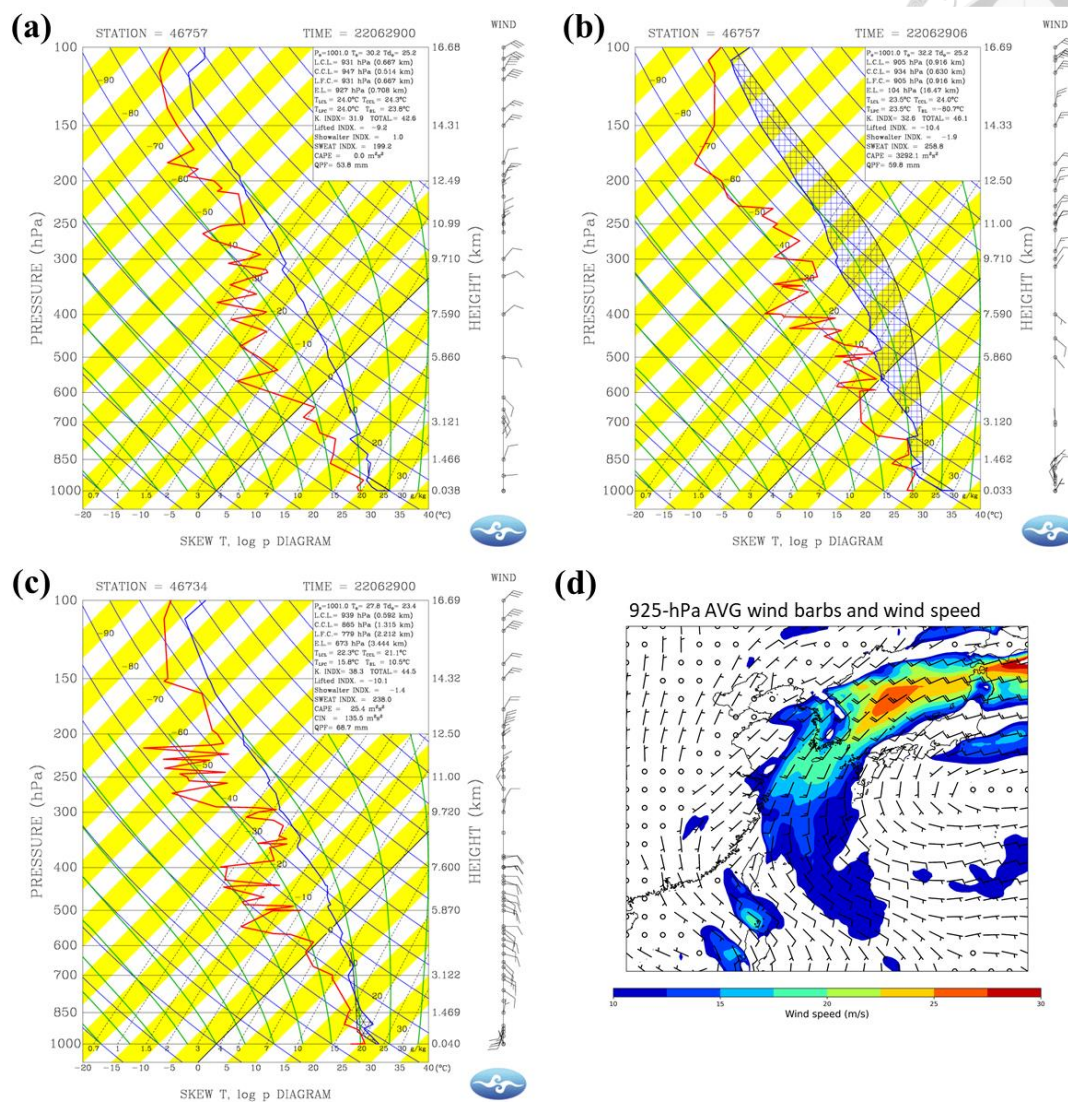
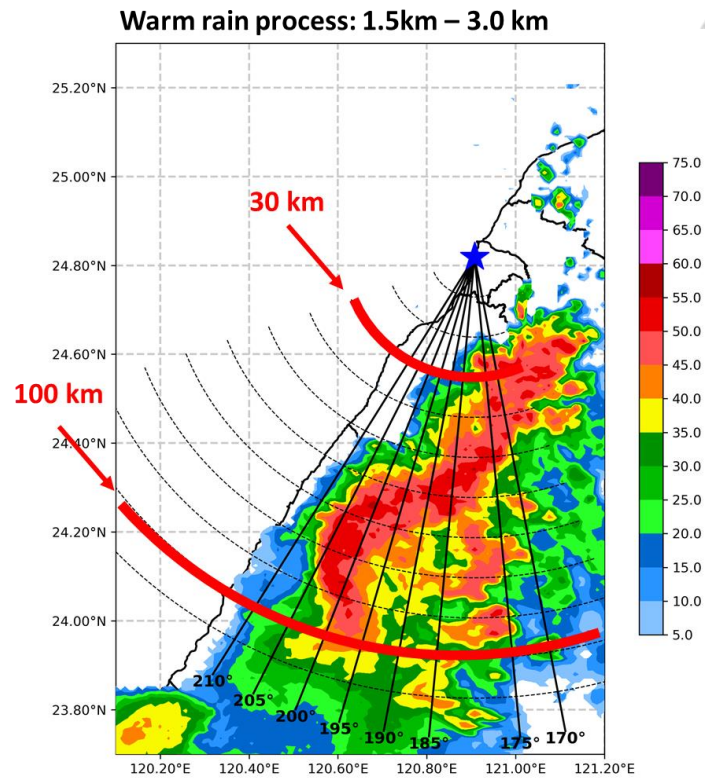


Figure 28. (a) Sounding profile from Hsinchu at 00 UTC, (b) sounding profile from Hsinchu at 06 UTC, (c) sounding profile from Makung at 00 UTC, and (d) 925-hPa wind barbs and wind speed (shading).

(a)



(b)

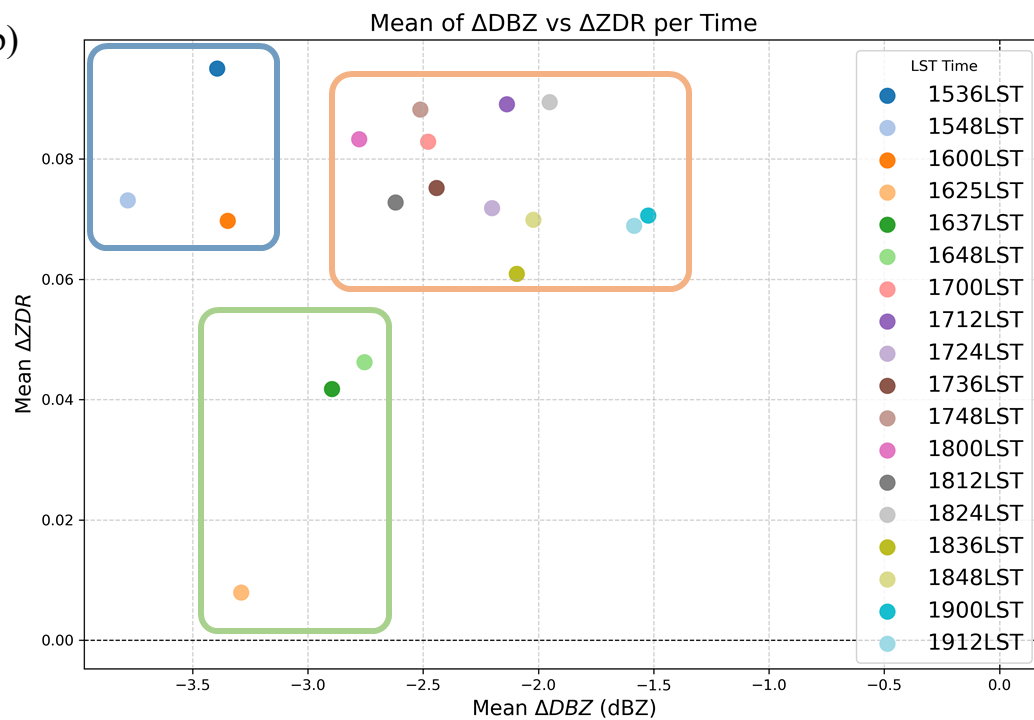


Figure 29. (a) Definition of the concentric horizontal analysis domain with radii from 30 to 100 km, centered on the S-Pol radar (blue star). The domain is based on the radar's observable altitude coverage from the gridded PPI reflectivity data. (b) Diagram adapted from Kumjian (2014; 2022) to illustrate warm-rain processes, showing only the second quadrant. Different line styles represent different time periods.

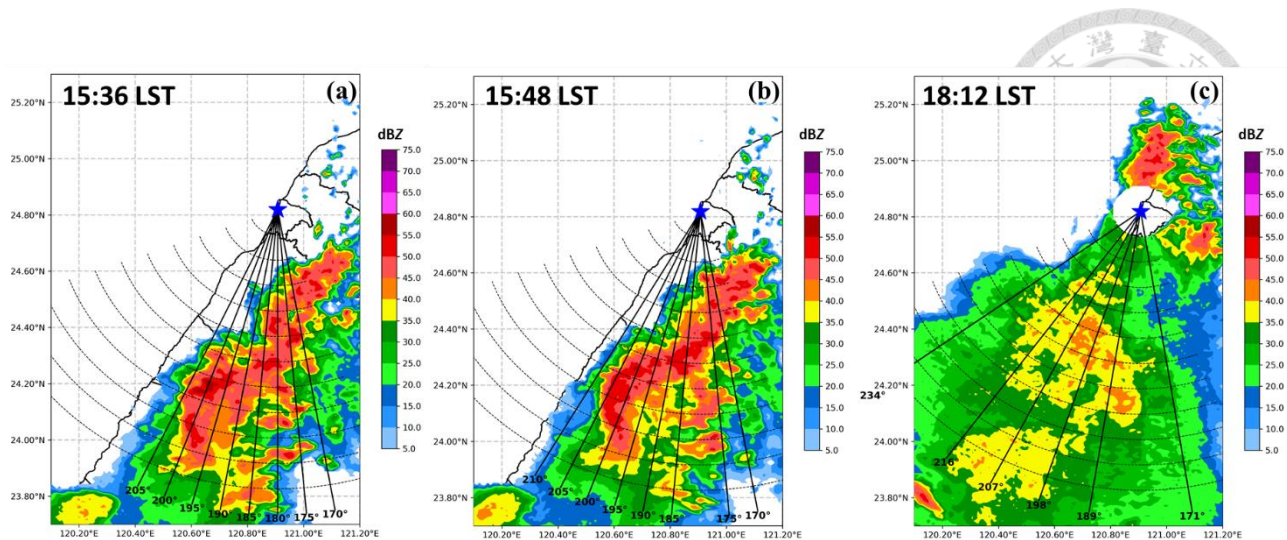


Figure 30. Locations of the convective system at 1536 and 1548 LST, along with RHI scans from various azimuthal angles. Concentric semicircles are drawn at 10-km radial intervals.

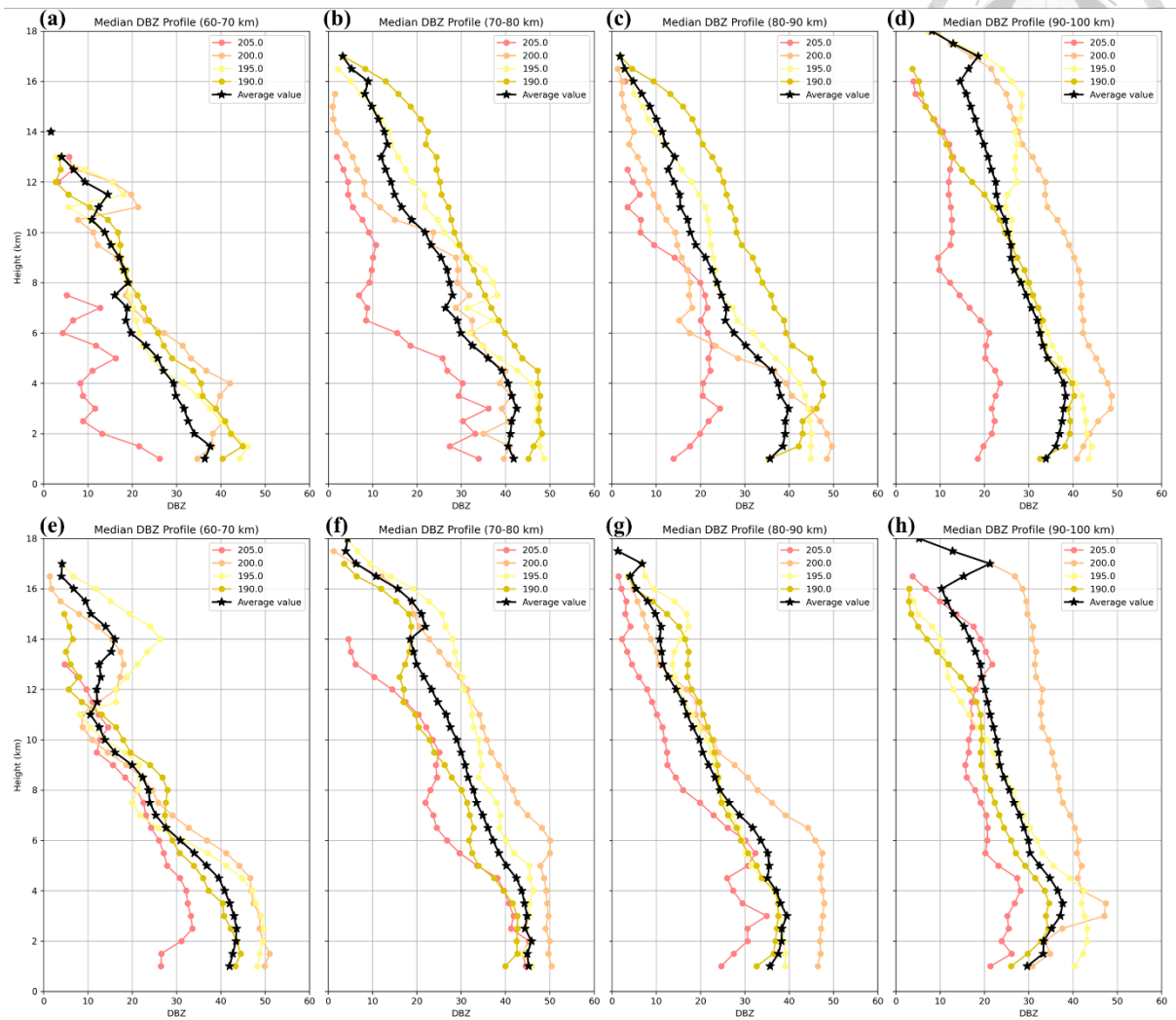


Figure 31. Median CFADs of radar reflectivity (dBZ) within the 60–90 km range for RHIs between 190° and 205°. The (a)-(d) correspond to 1536 LST, and (e)-(h) correspond to 1548 LST.

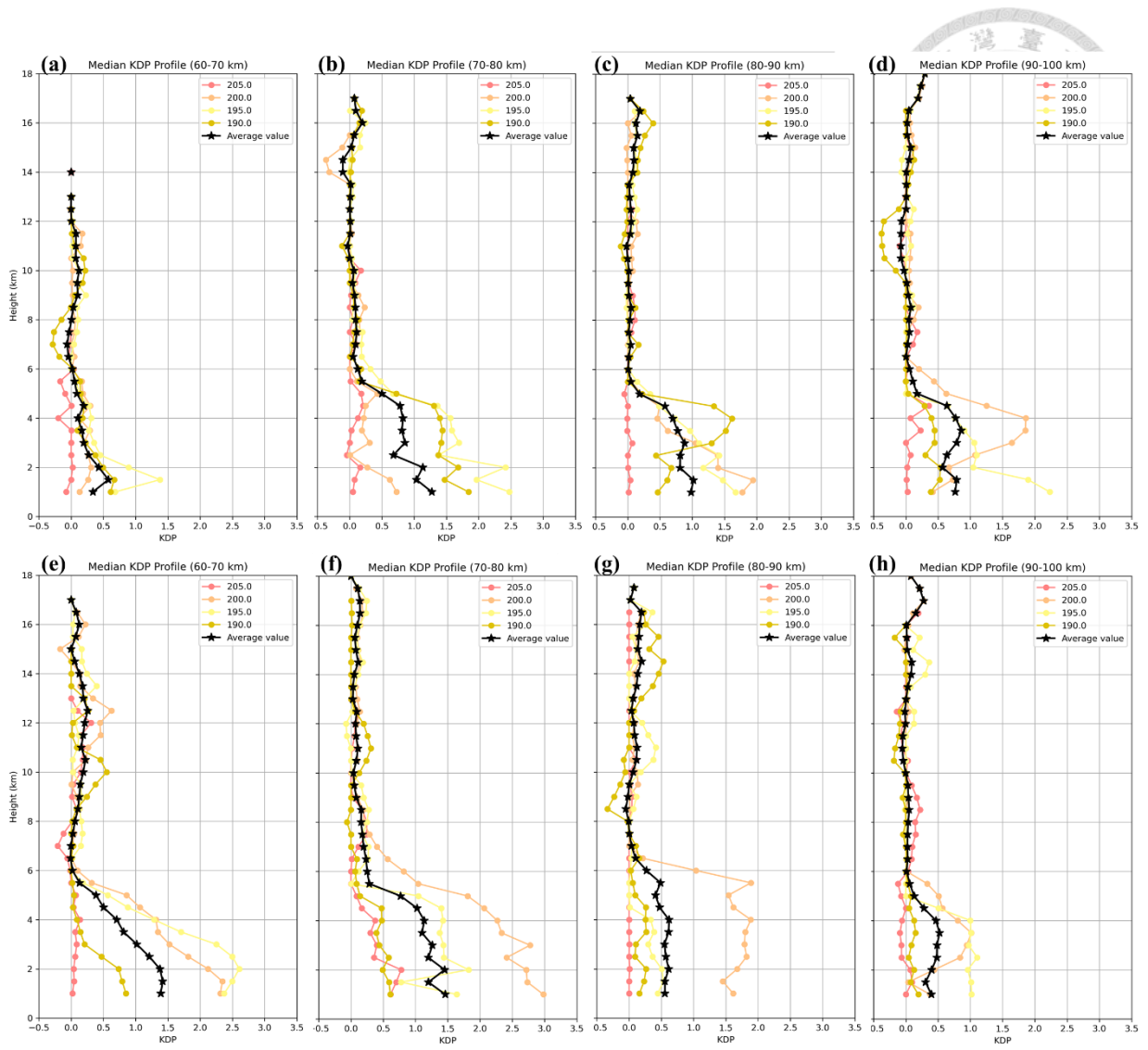


Figure 32. Same as Fig. 31, but for K_{DP} . (a)-(d) correspond to 1536 LST, and (e)-(h) correspond to 1548 LST.

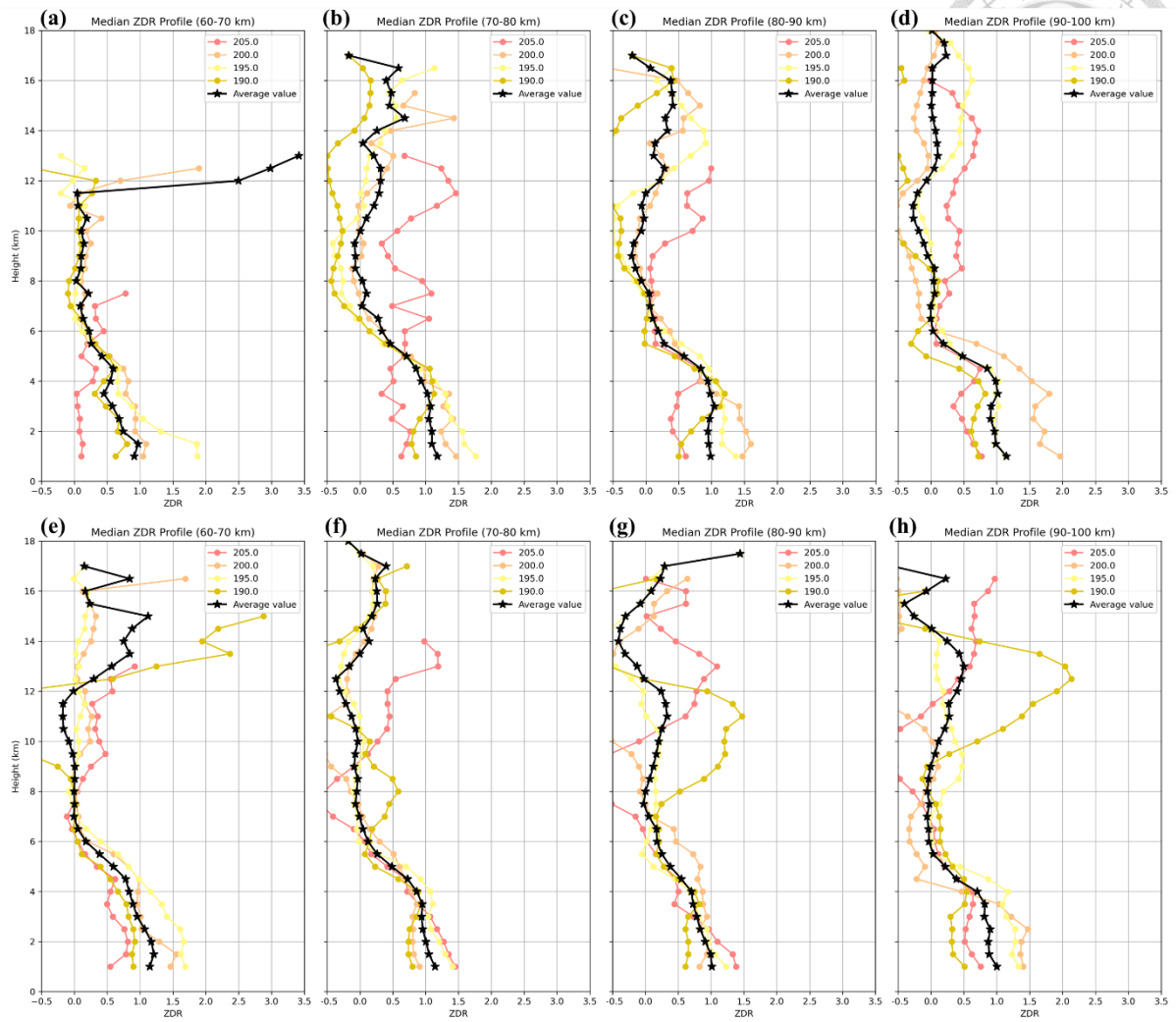


Figure 33. Same as Fig. 31, but for ZDR. (a)-(d) correspond to 1536 LST, and (e)-(h) correspond to 1548 LST.

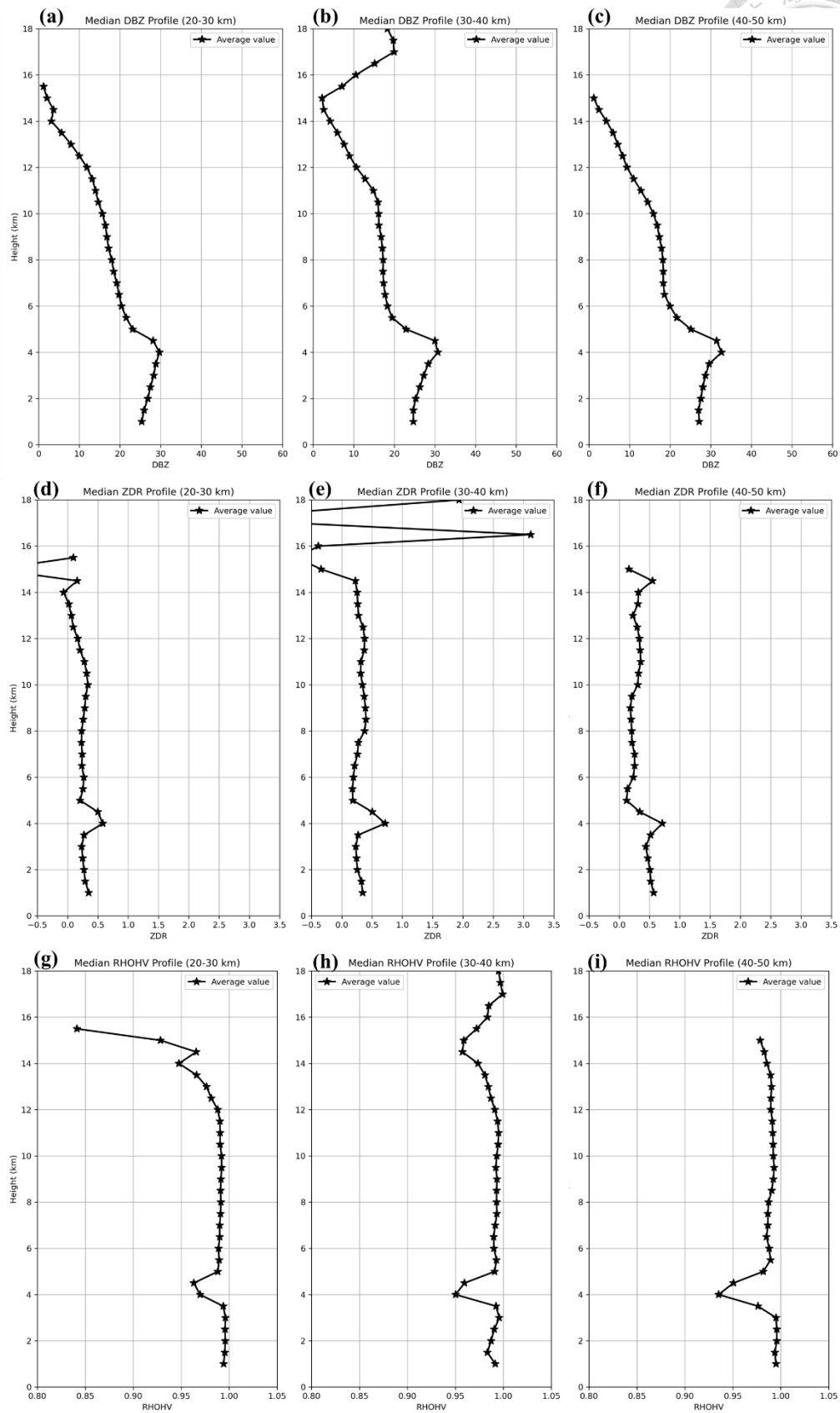


Figure 34. Azimuthal averages of RHIs during the stratiform precipitation period at 18:12 LST: (a)–(c) for DBZ, (d)–(e) for ZDR, and (f)–(i) for RHOHV.

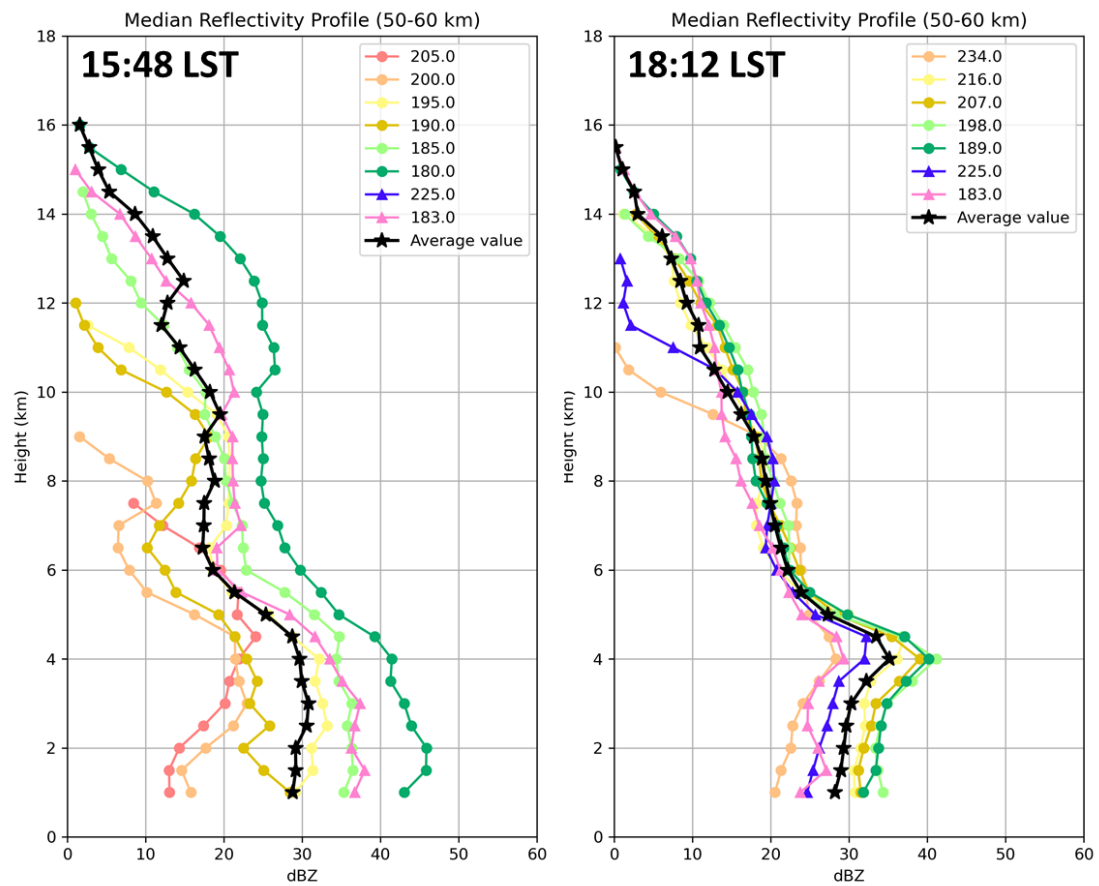


Figure 35. Same distance interval as in Fig. 34, illustrating distribution differences between the convective and stratiform regions. The RHI azimuths correspond to those shown in Fig. 30b and 30c. Different colored lines represent individual azimuthal directions, while the black star-marked dashed line indicates the mean profile. Panel (a) represents the mean profiles in the convective region at 1548 LST, and panel (b) represents the mean profiles in stratiform region at 1812 LST.

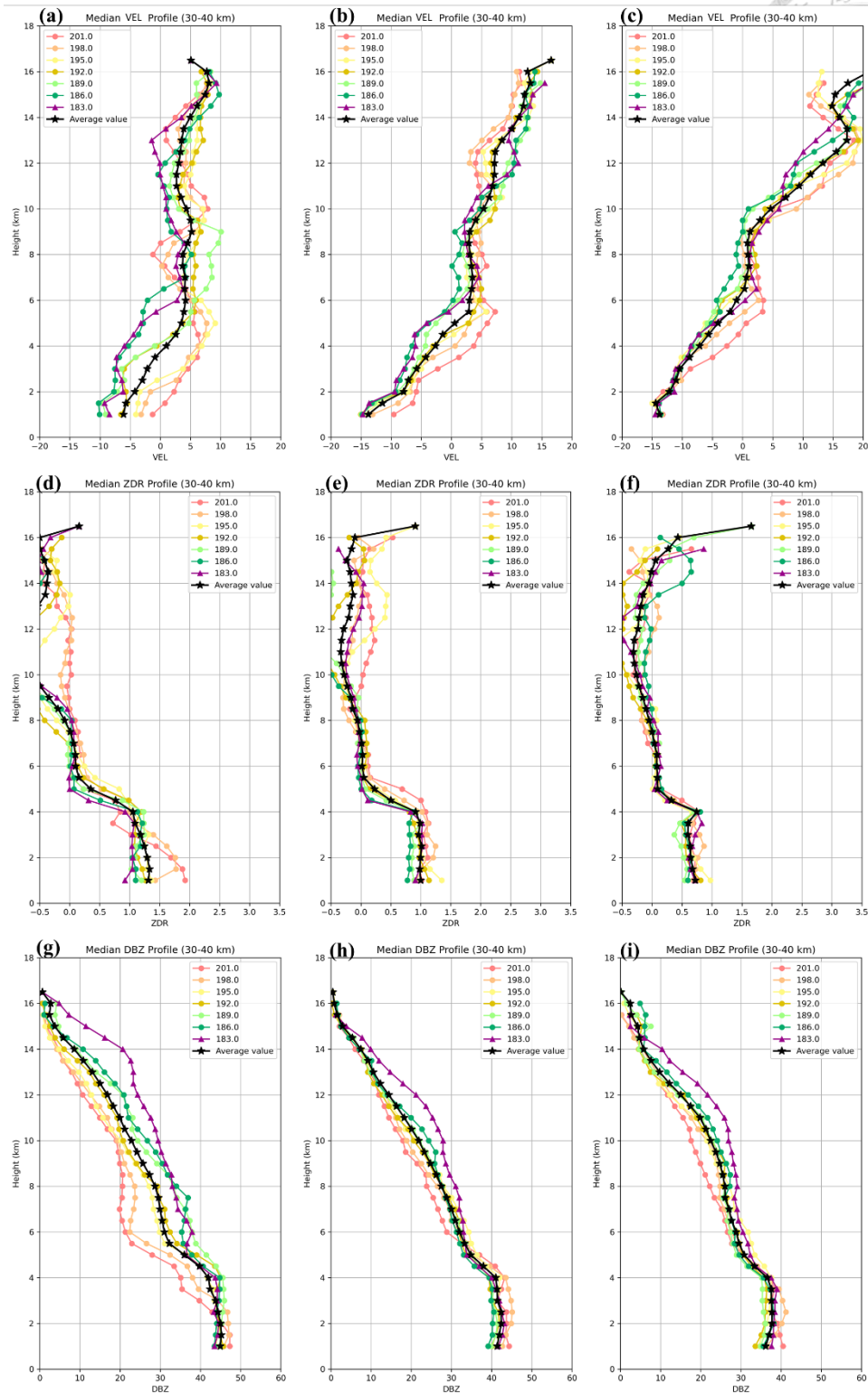


Figure 36. Differences observed in the 30–40 km range before and after the arrival of the RTF flow. Different colored lines represent individual azimuthal directions, and the black star-marked dashed line indicates the mean profile. The vertical panels show radial velocity, Z_{DR} , and Z_H in panels (a), (d), and (g), respectively. The horizontal panels represent different times: 1648, 1700, and 1712 LST in panels (a), (b), and (c),

respectively.

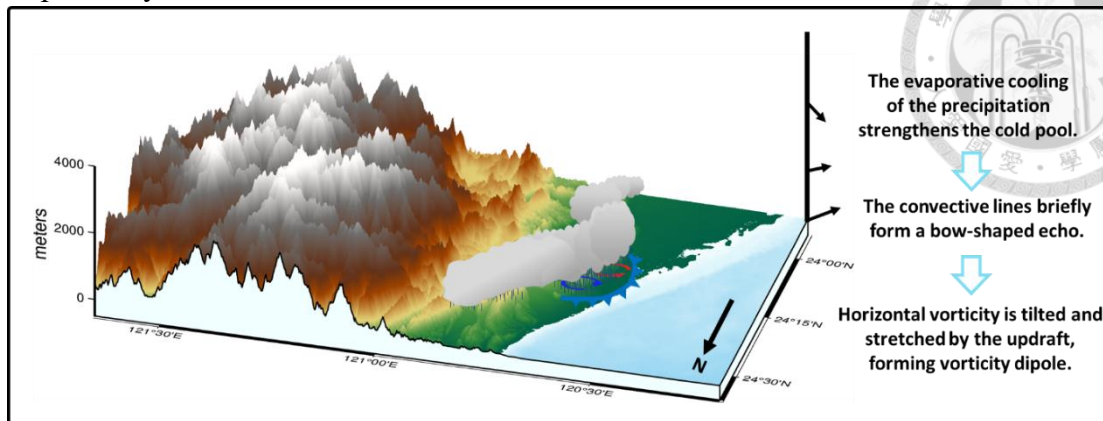


Figure 37. Conceptual diagram illustrating the formation of a bow-shaped echo.

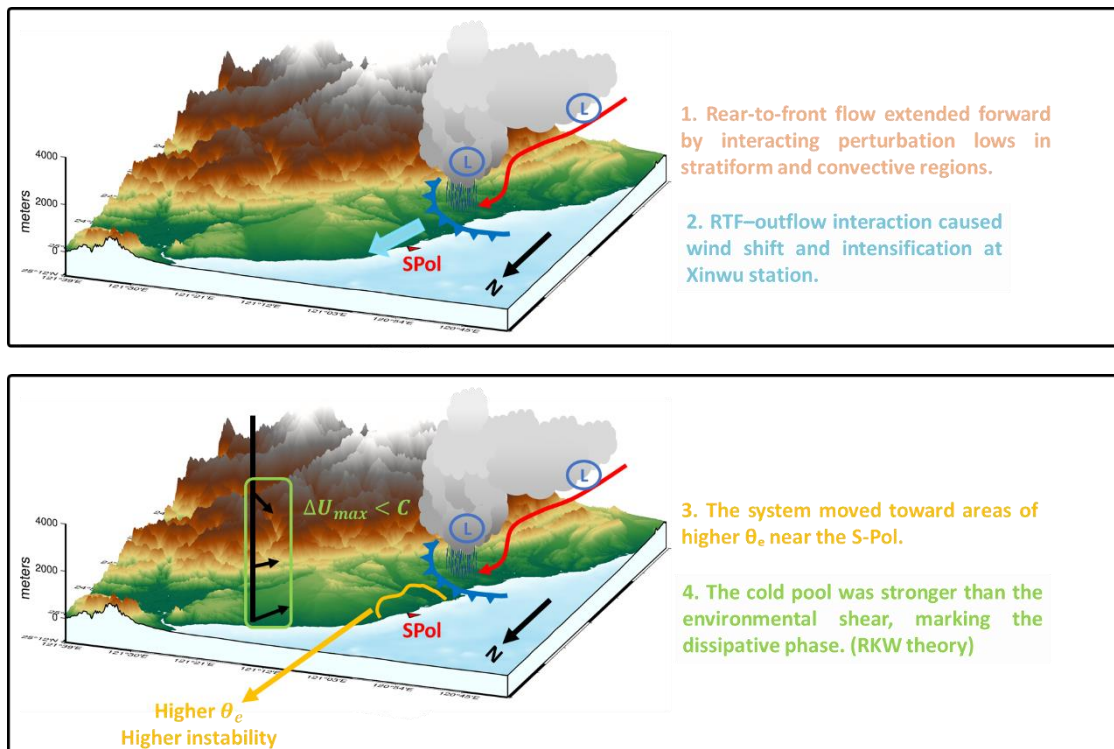


Figure 38. Schematic diagrams for the key processes during the northward propagation and dissipation of the QLCS.



Tables

Table A1. Convective initiation (CI) and non-CI events that met weak synoptic conditions and included additional sounding observations during TAHOPE.

CI	Non-CI
6/25, 8/1, 8/3, 8/4	7/12, 7/13, 7/14, 7/21, 7/22, 7/23, 7/24

Figures

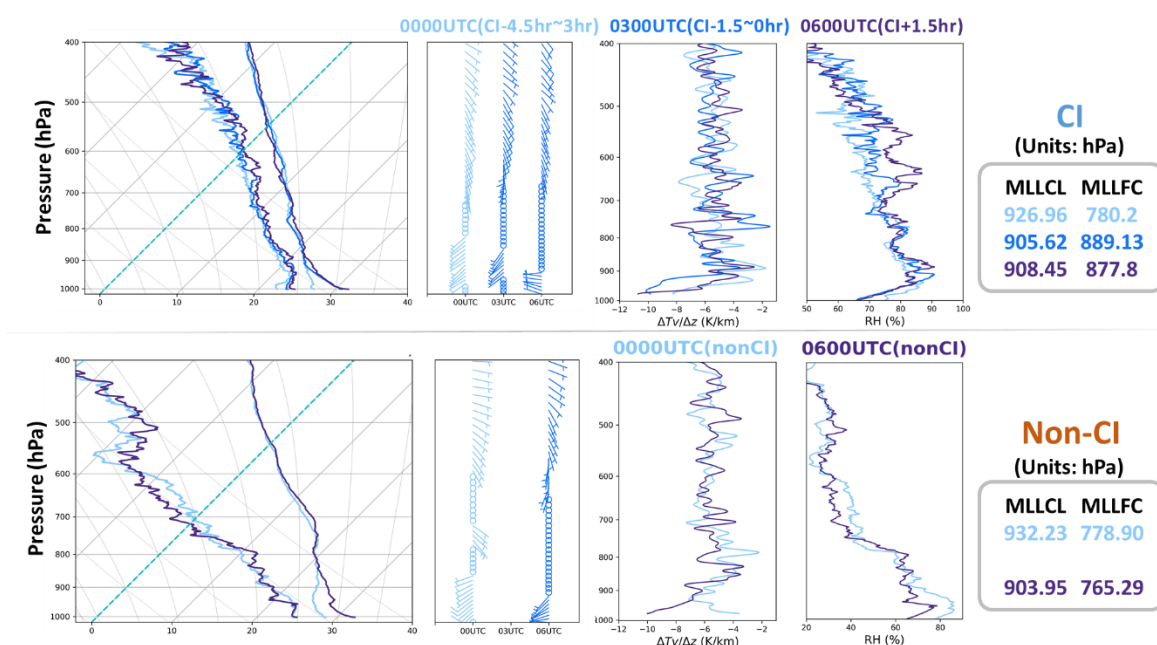


Figure A1. Composite analysis of the CI and non-CI events. The light to dark blue shades represent different sounding launch times (00, 03, and 06 UTC). From left to right, the panels show the sounding profiles, wind direction and speed, instability, and relative humidity. The top row depicts the composite average for CI events, while the bottom row shows the composite average for non-CI events.

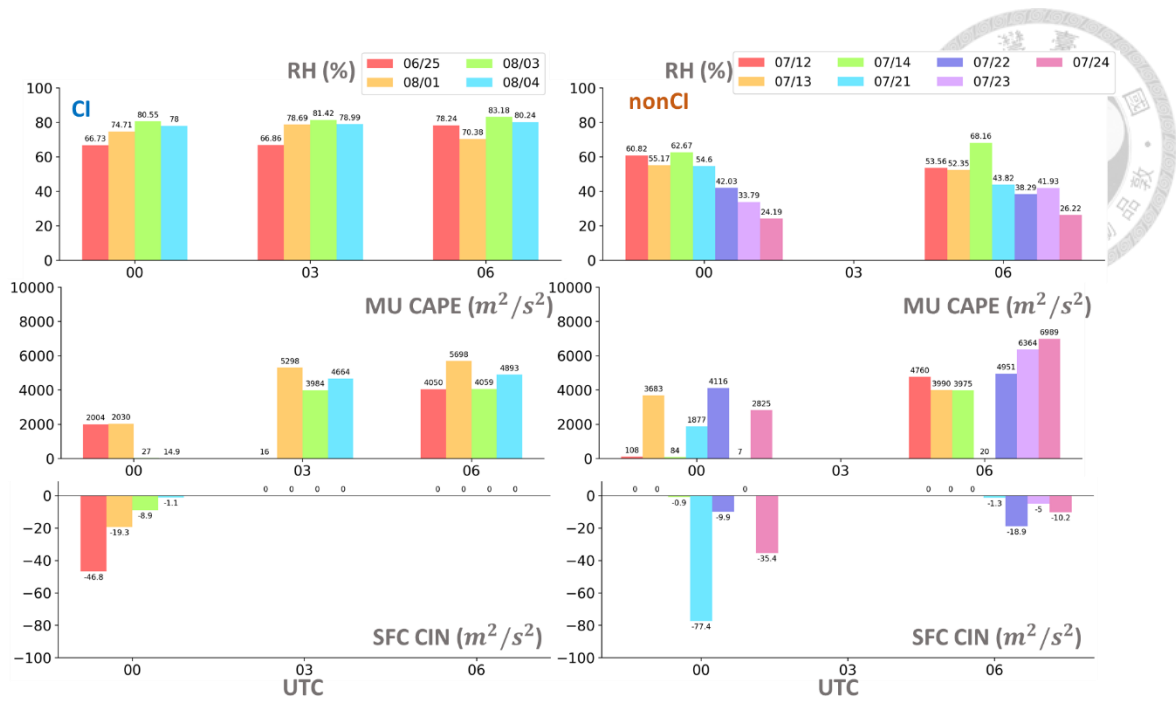


Figure A2. The left column represents the CI parameters, while the right column represents the non-CI parameters. Each colored bar indicates a different day, and the horizontal axis corresponds to the sounding launch time. Each horizontal row displays different variable (relative humidity, MU CAPE, and SFC CIN).

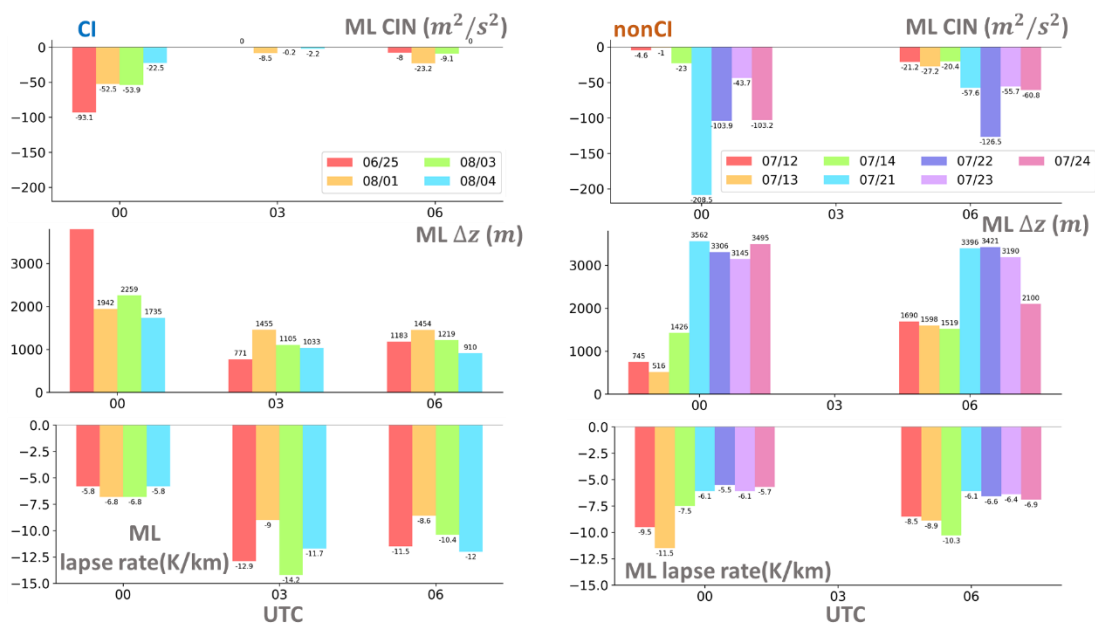


Figure A3. Same as Fig. B, but for different variables (ML CIN, ML Δz , ML lapse rate)

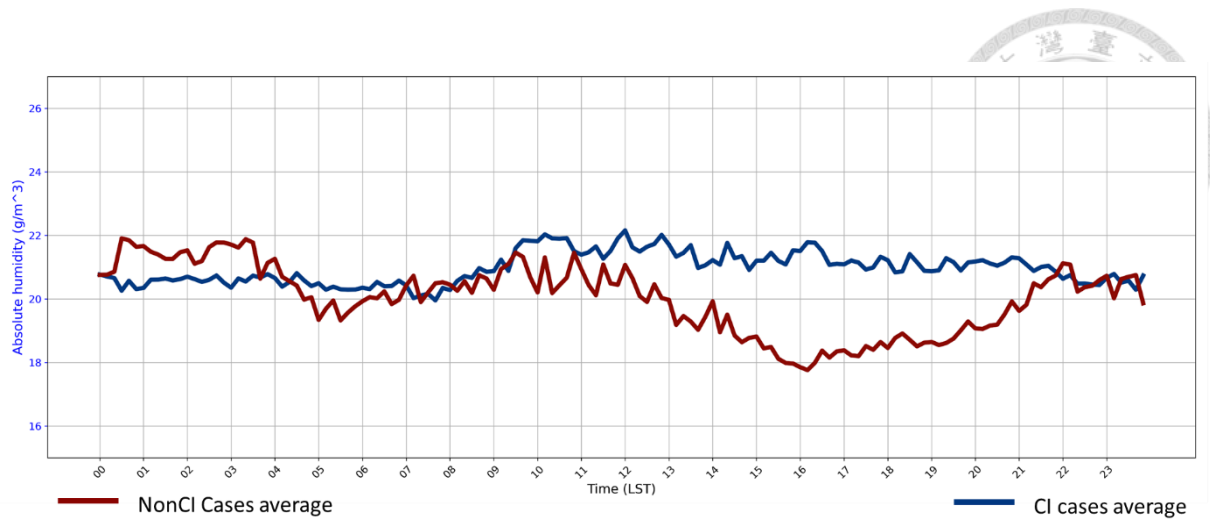


Figure A4. Mean absolute humidity at 521-m altitude for CI and non-CI cases observed by the MPD at Hsinchu station.