Vapor Isotope Probing of Typhoons Invading the Taiwan Region in 2016

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Abstract Every year, several typhoons originating from the West Pacific come close to Taiwan, modulating the regional weather through intense rainfall and enhanced humidity. The isotopic ratios (δ18O and δD) in the typhoon vapor evolve along its track due to rainout, recycling, and replenishment. In 2016, this signature was recorded in Taipei as typhoon vapors invaded the local atmosphere. After typhoon disintegration, the isotope ratios of the local atmosphere slowly return to the background value. The fall and rise in the ratios give rise to a V-shaped variation. The processes controlling the fall (like moisture turnover, typhoon speed, wind speed, and direction) are fast (1–2 days). In contrast, the recovery is slow (several days) depending on the replacement rate of the local atmospheric moisture by incoming oceanic vapor. An isotope evolution model based on Rayleigh condensation and a box model of moisture mixing in the recovery phase are proposed and employed to explain the V-shape. The combination of these two models for four major typhoons (Nepartak, Megi, Meranti, and Malakas) shows close agreement with observations. The evolution model suggests that the precipitation efficiency, PE (rainfall/overhead moisture), and typhoon size are two important parameters controlling the decrease in ratio. The characteristic turnover time of typhoon moisture (inverse of PE, obtained from European Centre for Medium-Range Weather Forecasts fifth Re-Analysis data) is estimated to be ~2 days for a Pacific typhoon of 500 km radius. Our isotope study provides important constraints on the moisture budget of Pacific typhoons.

Plain Language Summary Tropical typhoons transport heat and moisture from the ocean surface to the atmosphere at fast rates. To understand the moisture input and output during the typhoon evolution, stable oxygen and hydrogen isotope tracers in the vapor have been continuously measured at a station in Taipei, Taiwan. Four major typhoons originating in the western North Pacific in 2016 passed near the Taiwan region. The heavy-to-light isotope ratios of typhoon moisture that reached Taipei were analyzed. The isotope ratios of moisture changed with time due to rainout-related fractionation processes and subsequent mixing with atmospheric vapor producing a V-shaped pattern as the typhoon moved and disintegrated. An idealized model is presented to show how a Rayleigh fractionation operates through step-wise rainout from existing vapor mass that is supplemented by a continuous import of moisture by low-level inflow and ocean evaporation. The recovery of the isotope ratios to pre-typhoon values takes place by slow mixing of the incoming oceanic vapor with the existing atmospheric vapor over the station.

1. Introduction

Typhoons are serious natural hazards that sensitively respond to the rising SST due to global warming. Bender et al. (2010) proposed a hurricane-prediction model that projects nearly a doubling of the frequency of category 4 and 5 storms by the end of the 21st century. It is well-known that the water budget of tropical cyclones is intimately connected to their intensities (Emanuel, 2003; Ozawa & Shimokawa, 2015), but the detailed nature of moisture supply is still a debatable issue. For example, the mean rainfall in tropical cyclones is often an order of magnitude higher than the flux of evaporated oceanic moisture from the cyclone footprint. The surface evaporation within 500 km from the center contributes only 10%–15% of the rainfall (Vanniere et al., 2020). In case of North Atlantic hurricanes, the concurrent evaporation from a circular zone of about 3,000 km radius supplies only about 10%–20% of the total rainfall (Makarieva et al., 2017). Despite this meager supply, cyclones sustain their rainfall for several days (Vecchi & Knutson, 2008; Zhao et al., 2019). Although enough rainfall can presumably be sustained if the vapor is supplied from a much larger area of the ocean (within about 2,000 km from the cyclone center), the available pressure gradient is not sufficient to drive this moisture toward the center where the rainfall is maximum (Trenberth et al., 2007). Based on the above arguments, several researchers
have invoked other moisture supply mechanisms to explain the observed rainfall. Water vapor available in the surrounding environment can be swept up by the cyclone through its movement (if faster than the background wind) and internal circulation (Makarieva et al., 2017). Trenberth et al. (2003) state that for cyclones, air spirals in to the system from large distances and moisture can be transported over scales much larger than the areas of heavy rain.

1.1. Physical and Meteorological Description of Typhoons

Tropical cyclones or typhoons generally develop over ocean water whose surface temperature exceeds 26°C (Anthes, 1974; Emanuel, 2003) and the water is warmed up to a depth of 40–50 m. A typhoon can be considered as a mature storm system morphologically cylindrical and comprising of a strong cyclonic vortex, characterized by organized precipitation zones. It has three main features: the eye, the eye wall, and the spiral rain/cloud bands (see Figure S1 in Supporting Information S1; henceforth, “S” denotes Supporting Information). The initial fuel for a tropical cyclone is provided by water vapor and heat from the warm ocean to the overlying air (Anthes, 1974). As the warm, moist air rises, it expands and cools, and releases latent heat through condensation of water vapor and rain formation (Gray, 1968). The column of air in the core of the disturbance is warmed and moistened by this process (Houze, 2010).

The typhoon season for the West North Pacific Basin extends over 3 months: July, August, and September (Henny et al., 2021). Reported in Global Guide to Tropical Cyclone Forecasting by World Meteorological Organization (Neumann, 2017), on average (1979–2008), 49 tropical cyclones (having sustained wind speed >33 m/s) develop globally each year and 17 of them originate in the West Pacific Basin. Typhoons that form in this region are usually big in size, covering a few thousand kilometers due to the large amount of moisture from the Pacific warm pool. Taiwan lies near the middle part of the band of tracks of west North Pacific typhoons, revealed by analysis of past trajectories (Rumpf et al., 2007). Consequently, this region experiences typhoon-associated intense rains several times a year. On average, the typhoon-associated precipitation is as much as 10% of the total annual rainfall in Taiwan (Henny et al., 2021). In addition to the well-identified typhoons, this region also experiences severe storm systems (Figure 1) which could be as intense but not as centrally organized as typhoons.

1.2. Isotopic Perspective of Typhoons

Stable oxygen and hydrogen isotope tracers (expressed in terms of delta-values, i.e., δ^{18}O and δD, where δ = R_{sample}/(R_{standard} − 1), and R = ^{18}O/^{16}O or D/H ratio of sample and standard) in the rain and atmospheric vapor provide an understanding of the source(s) of the in situ moisture, its transport, and dynamical evolution (Dansgaard, 1964; Gat, 1996). The equilibrium vapor over the ocean has ~10‰ lower δ^{18}O compared to the sea water. During the ascent of the vapor in the atmosphere, cooling, and rain formation, rainwater with higher δ^{18}O is produced and the δ^{18}O of the left-over vapor decreases (Rayleigh condensation). As condensation proceeds, the isotope ratios of the remaining moisture and the ensuing precipitation decrease (Dansgaard, 1964). Water vapors inside a tropical cyclone have oxygen isotope ratios that range from ~9‰ to ~26‰ (Gedzelman & Arnold, 1994). The lower isotope ratios represent vapors that have been processed several times by continuous rainfall. This shows that both vapor and rainwater can be used as tracers to quantify the inflow and outflow of moisture inside a typhoon, hurricane, or cyclone (Munksgaard et al., 2015; Sánchez-Murillo et al., 2019). Most earlier isotope tracing studies were done on single typhoons invading a particular location. To study multiple typhoons, the Taiwan region in the west Pacific provides an ideal place because it frequently experiences several typhoons in a year. The year 2016, in this respect, was a good period for studying a series of strong typhoons and storms that invaded this region (Figures 1 and 2).

The isotope analysis of rainwaters and groundwaters, specifically in East Asia organized by the IAEA (International Atomic Energy Agency), led to the understanding of their variations in terms of three major monsoon systems (Aggarwal et al., 2004). However, this work did not attempt to analyze the typhoon dynamics. Meteorological data and National Centers for Environmental Prediction (NCEP) re-analysis results show that Indian Ocean moisture dominates this region during the summer JJAS (June-July-August-September, southwest monsoon months). Low-level easterlies from the tropical western Pacific also contribute significantly during September (Aggarwal et al., 2004). Post-JJAS, winds from north China, Korean peninsula, and Japan carry moisture for the winter
Figure 1.
monsoon precipitation (Lim et al., 2002; Rangarajan et al., 2017). The average wind directions toward Taiwan in different months are shown in Figure S2 in Supporting Information S1, indicating diverse source regions. Thus, even in non-typhoon times, the vapors passing over Taiwan in different seasons have different isotope ratios.

During the typhoon, upper level air is brought down to the sea surface by downdraft and after mixing with the lower level air the joint air mass moves inward (Figure S1 in Supporting Information S1). This process is responsible for recycling of the moisture in a tropical cyclone. Lawrence and Gedzelman (1996) showed that the stable isotope ratios of rains in tropical cyclones are far lower than those in other tropical events and generally display a pattern of decrease toward the center. In their study, extremely low values were observed near the eye wall. The data were explained by a simulation model governed by fractionation during change of phase in rain formation and by diffusive isotope exchange between vapor and raindrops during their fall. In their model, as the air is uplifted, the moisture in it condenses to form clouds of water droplets that are enriched relative to the vapor. If the heavier condensates leave the cloud and fall down as raindrops, they may partially evaporate and exchange with the surrounding vapor at lower level which further increases their isotope ratios and depletes the vapor. Under intense updraft, the first-generation of depleted vapor is again lifted and gives rise to the second and third generations of rains that are progressively depleted. Thus an inward decrease of isotope ratios can be generated by step-wise isotope exchange between falling raindrops and converging water vapor. However, it fails to explain an inward increase in ratios found in high-altitude observations. In Hurricane “Faith” isotope ratios of rain and vapor at heights between 600 and 3,000 m over the ocean first decreased radially inward (up to about 100 and 300 km from the center) but then increased in the eyewall to values almost as high as outside the hurricane (Ehhalt & Ostlund, 1970). In case of Hurricane “Olivia,” isotope ratios of rain increased inward to the eyewall (Lawrence...
et al., 2002). Gedzelman et al. (2003) analyzed precipitation and water vapor collected from four hurricanes at flight levels between 850 and 475 hPa. The lowest isotope ratios were found in or near regions of stratiform precipitation that occurred between 50 and 250 km from the hurricane center. Surprisingly, the isotope ratios, in this case, increased inward from the eyewall.

In a study of water vapor and precipitation during the passage of the typhoon “Shanshan” over a small island near Taiwan (about 200 km away), Fudeyasu et al. (2008) found that the isotope ratios decreased radially inward in the typhoon's outer region but anomalously high isotope ratios appeared in the inner region. A broad echo indicating stratiform precipitation was found at the outer rain shield while two inner rainbands indicated an intense convective zone. They explained the observed isotopic features by a high rainout effect due to increased condensation efficiency. However, in the inner-most region, the water vapor was isotopically enriched by exchange with sea spray and sea surface vapor which have heavy isotope ratios.

There have been a few precipitation isotope studies in Taiwan (Peng et al., 2010, 2011); two investigations were done for the isotope distributions in rain and vapor phases (Laskar et al., 2014; Rangarajan et al., 2017). Laskar et al. (2014) showed that the isotopic composition of water vapor varies between $-11\%_{\text{e}}$ and $-21\%_{\text{e}}$ for $\delta^{18}O$ and $-75\%_{\text{e}}$ and $-150\%_{\text{e}}$ for $\deltaD$, but no clear seasonal variation was observed (for the years 2011–2013). The average difference between rain and vapor $\delta^{18}O$ was $8.6 \pm 1.1\%_{\text{e}}$. The d-excess (defined as $d$-excess $= \deltaD - 8 \times \delta^{18}O$; Dansgaard, 1964) in rainwater did not vary significantly and was on average, $9.6\%_{\text{e}}$, a value slightly higher than that of the vapor which had large variations with a mean of $7.3\%_{\text{e}}$. To study how rainwater interacts with water vapor, two events were selected: a Mei-Yu event (June 12–16 in 2012) and a typhoon named Saola (31 July to 4 August in 2012). They found that the vapor and the rainwater were nearly in isotopic equilibrium during typhoon while they deviated significantly during the Mei-Yu event. Rangarajan et al. (2017) studied the winter time vapor source (during 2011 and 2012) by analyzing high-resolution d-excess variations. They developed a box model to show that both high and low d-excess events in the winter are primarily controlled by the humidity deficit over the ocean over which the air mass moved.

### 1.3. Paleo-Records of Typhoons

The vapor, rain, and concentration data of this study show that the whole of eastern China and Taiwan regions are flooded with low isotope ratio moisture in specific seasons which have important implication in the regional isotope hydrology models. Because the intense rains associated with typhoons and storms have low ratios, they can act as discrete signals in isotope hydrometeorological investigations (Skrzypek et al., 2019). Such low values are known to be preserved in palaeo-records like tree rings and stalagmites (Frappier et al., 2007; Miller et al., 2006). Interpretation of isotope data in such geological systems, however, requires a better understanding of the dynamics of water and moisture recycling in modern typhoons and their attendant effects on isotopes.

### 1.4. Motivation for the Study

As mentioned earlier, fresh evaporation cannot supply enough vapor to sustain the intense rainfall associated with a typhoon. Makarieva et al. (2017) proposed that as the cyclone moves along a track, it scavenges the required moisture from the sectional area of encounter. During the rainfall over the footprint zone, the cyclone depletes the pre-existing vapor in its wake. In addition, a large component of the typhoon moisture is recycled. The principal motivation of the present work was to see if this conjecture can be checked quantitatively by a vapor isotope study.

We report high-resolution $\delta^{18}O$ and d-excess values of vapors of typhoons and storms that originated in the western Pacific and came close to Taiwan in the year 2016 before landfall in Taiwan, China, or Japan. We also report the rain isotope data during the studied typhoons. However, the rains were discontinuous and sparse, and we cannot use these data for moisture budget studies. Instead, we use the vapor isotope time series data to decipher the amount of moisture recycling inside a typhoon and fresh moisture input. Our calculations are based on several parameters (like moisture contributing area, the total overhead moisture, rainfall along the track, etc.) and are described in detail in Section 4.3.
2. Experimental Setup

The atmospheric water vapor and rainwater as well as moisture were collected at a station in northern Taiwan (rooftop of Institute of Earth Sciences, Academia Sinica, Taipei: 25.0424°N, 121.6161°E, approximately 25 km from the coast; Figure S3 in Supporting Information S1). The isotope ratios and moisture concentrations were analyzed using a cavity ring-down spectrometer (CRDS Picarro L2120-I, USA; Crosson, 2008; Gupta et al., 2009). For water vapor, the air was taken in using an air blower and introduced to the spectrometer through an inlet at an elevation ~50 m above the ground (Figure S3 in Supporting Information S1). All isotopic ratios are reported in δ notation (in ‰) relative to V-SMOW (Vienna Standard Mean Ocean Water) with analytical uncertainty of 0.1‰ in δ¹⁸O and 0.8‰ in δD values and ~100 ppmv in concentration (for details see Laskar et al., 2014). The instrument was calibrated regularly (1 week to 10 days) with six working water standards, calibrated against VSMOW2 and SLAP2: δ¹⁸O and δD values are: −17.88, −138.09‰; −15.93, −116.64‰; −13.37, −98.83‰; −11.28, −82.02‰; −6.98, −54.79‰; and −6.44, −35.08‰ for water vapor and slightly heavier (ranging −15.36‰ to −1.40‰ for δ¹⁸O and −113.56‰ to −10.44‰ for δD) for rainwater. The deviation from an expected 1:1 line between the measured and actual values of the standards was used to correct the measured values of the samples. No significant variation in linearity is observed in the corrected isotopic composition for water vapor concentration varying from 10,000 to ~35,000 ppmv. Though the temporal resolution of measurement was high (about 2 s), the data set was hourly averaged for presentation and subsequent analysis and interpretation. Because of the high level of suspended particles (the concentration of PM₁₀ ranges between ~10 to >50 μg/m³) in the region, a particle filter is installed in the air inlet (Figure S3 in Supporting Information S1). The filter holder has a volume of ~50 ml, due to which the resolution less than ~2 min is lost (the flow to the analyzer is ~30 ml/min).

We present the δ¹⁸O, δD, and d-excess values of vapors along with their concentrations, each number representing average for a period of 24 hr for each date (midnight to midnight). The isotope data are available at Harvard Dataverse [https://doi.org/10.7910/DVN/UTE3Q1](https://doi.org/10.7910/DVN/UTE3Q1) (2018; Wei et al., 2019; Table S1 in Data Set S1). The local meteorological parameters like wind speeds, wind directions, and pressures (station code: 466920), along with satellite images and radar maps, were obtained from the Central Weather Bureau of Taiwan (CWB), archived by Taiwan Data Bank for Atmospheric & Hydrologic Research. In addition, we used the European Centre for Medium-Range Weather Forecasts (ECMWF) fifth Re-Analysis (ERA5) data set (Hersbach et al., 2020). A brief description of ERA5 is given in the Supporting Information.

3. Results

The variations of δ¹⁸O and d-excess of the vapor (measured in Academia Sinica, Taipei and averaged over 1 day), the vapor concentrations and daily rainfall (taken from a nearby Meteorological station) during 2016 are plotted in Figure 1. The gaps in the data correspond to the periods when there was temporary breakdown of the CRDS Picarro system or the standard values were not acceptably accurate. In 2016, we had reliable data for a total of 267 days. The average (daily) vapor isotope ratios δ¹⁸O, δD, and d-excess measured in Taipei range from −10.9 to −24.7, −74‰ to −185‰, and 3.7‰ to 35.7‰, respectively (Figure 1 shows data for δ¹⁸O and d-excess). The grand average values (including the rapid oscillations) are −14.4, −100.5, and 14.9‰, respectively. The corresponding ranges and average values for the rains are −10.7 to +0.16, −81.04 to +0.75, −6.61 to +14.28‰, and −2.65, −10.20, 11.00‰, respectively. The average difference between the δ¹⁸O values of rain and vapor is 12.1 ± 1.3‰. The vapor concentrations (in ppm) range from a minimum of 10,200 to a maximum of 36,600 with an average of 26,000. The winter season is characterized by low values of concentrations whereas the summer season (coinciding partly with the typhoon season) by higher values. The winter vapor has fluctuating d-excess values with some of the values being very high (as much as 36‰). In contrast, the summer has consistently lower values within the range 5‰ to 15‰. As expected, the storms and typhoons have profiles of negative excursions (decrease followed by increase) of isotope ratios (δ¹⁸O and δD) displaying typical V-shaped variations (insets in Figure 1).

3.1. V-Shaped Variations in the δ¹⁸O and δD Values

The δ¹⁸O values have an upper limit of about −12‰, representing oceanic vapor near the surface (Good et al., 2015). The oceanic steady values were superposed with lower δ¹⁸O values whenever organized meteorological systems
(typhoons, storms, and tropical depressions) denoted by arrows in Figure 1 invaded the region. The typhoons or storm/depression systems exhibit rapid excursions of the isotopic ratios to highly negative values followed by recovery to pre-typhoon values (in the form of approximate V-shaped spikes; Figure 1). Such large negative vapor values are essentially caused by Rayleigh Fractionation induced by the intense rainfall associated with them. The negative peaks of the δ\(^{18}\)O values during such events range from −15.4‰ (storm Ambo) to −20.5‰ (typhoon Meranti) and last from a few days to about 12 days (typhoon Nepartak). Corresponding δD values range from −108‰ to −190‰. Eleven such systems were identified, some of which were named (indicated in Figure 1) by the regional countries. Out of these, four major typhoons (≥category 4), namely, Nepartak, Malakas, Meranti, and Megi, are selected here for investigation in detail (see Table S1 in Supporting Information S1 for information of these typhoons). Some of the depressions and storms, namely, Aere and Tokage, have highly depleted δ\(^{18}\)O values with V-shaped profile similar to typhoons. However, owing to the lack of any organized cloud structure that spatiotemporally evolves over a fixed path (like typhoons), it was difficult to model the vapor isotope evolution in these depressions and storms. The tracks of the abovementioned four typhoons (adopted from typhoon tracker of Japan; Figure 2) show that all of them originated in the western North Pacific warm pool where the surface waters are 2–5°C higher than other equatorial waters (Yan et al., 1992). The low-value excursions of δ\(^{18}\)O and δD run parallel to each other but their ratios are not constant and this shows up in the deviations of the d-excess due to reasons discussed below. The limited data shows that the depletion in rain δ\(^{18}\)O values in typhoon Nepartak, Meranti and Malaks coincide with those of the corresponding vapors.

### 3.2. Dependence of d-excess Values on Wind Direction

It is noticed that the d-excess variation is quite large (from 5‰ to 40‰) being largely determined by the season. During the winter months of January-February-March, the average values are 21.2 ± 7.1‰ (1σ standard deviation; with maximum and minimum values being 38.3‰ and 5.0‰, respectively). The months of November-December also have high d-excess values (on average 20.5 ± 5.7‰ with maximum and minimum values being 32.0‰ and 6.4‰, respectively). Unlike d-excess, the δ\(^{18}\)O values do not show significant seasonal variation except during the typhoon season when there are large negative excursions. There seems to be a prolonged period of lower δ\(^{18}\)O values during the typhoon season. This is because the measuring station is located in a basin that prevents quick full-scale freshening of air with the oceanic vapor. During the typhoon period, seven depressions/typhoons inundated the Taiwan region with negative isotope ratio vapors one after another (Figure 1). Therefore, the station did not have sufficient time to get back to the oceanic value of −12‰ after disintegration of one typhoon. This set of typhoon events created a general depression of the δ\(^{18}\)O in the August-October period. Broadly speaking, the average d-excess values during the months of December-January-February are usually high; and among them, there are many abnormally large values going as much as +40‰. Surprisingly, this part of the season also has very small values, which are associated with change in wind direction (Rangarajan et al., 2017).

The high values have been explained by the arrival of large vapor mass derived from East China Sea or Yellow Sea by dry continental air (coming from north China-Korea-Siberia). Kurita (2013) pointed out that high d-excess values result whenever evaporation occurs under a large humidity deficit and a strong temperature contrast between the surface air temperature and sea surface temperature (Gat et al., 2003; Uemura et al., 2008). In the northwestern Pacific, during the winter monsoon, cold and dry continental air mass flows over the relatively warmer seas, resulting in enhancement of evaporation involving kinetic fractionation, which depletes the \(^{18}\)O/\(^{16}\)O ratio proportionately more than the D/H ratio (Rangarajan et al., 2017). Therefore, vapor associated with these air masses, especially those with high wind speed, have higher d-excess values. In support of this hypothesis, we find a distinct correlation (Figure 3) between the d-excess values and the wind direction (wind vector field is shown in Figure S2 in Supporting Information S1). The maximum effect is seen in the January-February vapors. The winds coming from eastern seas carry normal oceanic vapor with d-excess of about 10‰–12‰ whereas winds from the north or northeast carry high-value vapors. Therefore, except for a few occasions, the high values are absent in the summer season when the air mass is mainly from the south or southeast or east (summer monsoon).

As mentioned, the d-excess could occasionally be lower even during the winter months, which is due to changes in the wind direction toward easterly or southerly. Two cases are shown in Figures 3a and 3b; the first one occurred during the fourth week of January and the second one during the second week of February 2016. The d-excess changed from a high value of around 40‰ on January 25–26, 2016 to a low value of about 15‰ on January 27–28. The wind is seen to have changed correspondingly from northeasterly (from 50°) to easterly
(90°) during this period (Figure S2 in Supporting Information S1). The anti-correlation is statistically significant (insets in Figure 3) although not perfect. This is because the apparent wind direction in Taipei (which we used here) does not precisely tell its origin, and in January, there are swings in direction as seen in the regional wind field diagram (Figure S2 in Supporting Information S1). Rangarajan et al. (2017) discussed in detail several cases of such swings in d-excess that are correlated with wind direction changes and other parameters in the winters of 2011 and 2012. They found that the wind direction change could explain about 60% of the variance.

3.3. Typhoon Nepartak

Nepartak formed as a tropical storm south of Guam on July 3, and began moving northwestward on July 4 (Figure 2); it acquired sufficient wind strength to be designated as a typhoon on the next day. Nepartak reached peak intensity with a pinhole eye on July 6 and had a peak windspeed of 280 km/hr with an estimated central pressure of 900 mb. It crossed Taiwan before emerging into the Taiwan Strait on July 9 in a weakened form, finally making landfall in China on July 10 as a tropical storm. The rainfall over Taipei throughout the traverse of Nepartak occurred from the outer-rain bands and possibly from stratiform clouds.

During the movement, the position of Nepartak eye changed rapidly with time (Figure 2). Figure 4 shows the cloud top infrared images showing the positions of the eyes of the four studied typhoons with respect to the measuring station (denoted by N1–N5 for Nepartak; Figure 2) as identified from satellite infrared pictures. The closest
approach of Nepartak to the station is on July 9 but the isotope ratio was the lowest on July 10 night and early hours of July 11. By this time, the typhoon was disintegrating in coastal China (N6 and N7; Figures 4 and 5). The $\delta^{18}$O values were about $-13\%e$ on July 7 when the vapors were coming from the outer rain bands (Figure 4) and mixing with the earlier oceanic vapor; the eye was about 700 km from Taipei. Between July 7 and the landfall day on July 8, the $\delta^{18}$O decreased rapidly to a value of about $-17.5\%e$. By this time, the vapor was derived mostly from the outermost convection cell until 9th evening (N2–N5; Figure 4). The source of the vapor was from the same outer band, between 8th afternoon (13:30 hr) and 9th evening (19:30 hr), reflected in the nearly constant value at $-17.4\%e$. On July 10 (N6), the contribution from the vapor under the stratiform area added to that from the vapor from the convective area, which shows up as a second decrease to a value of about $-18\%e$. Subsequently, the vapor mass was completely replaced by the stratiform vapor on 11th (N7; Figure 4) which decreased the value to about $-19\%e$ (Figure 5).

The meteorological parameters measured at Taipei station during the period July 1–19 are shown in Figure 5. Note that the minimum pressure and maximum wind speed were between July 8 and 9, 2016 but the $\delta^{18}$O minimum at the Taipei station occurred ∼2 days later. This shows that the $\delta^{18}$O distribution inside a moving typhoon vortex is not uniform and the $\delta^{18}$O minimum is not coincident with the eye position but displaced toward the posterior part. Such a feature was noted by Fudeyasu et al. (2008) for the typhoon “Shanshan” who stated that “In the cyclone's outer region, the water vapor was isotopically depleted” and “In contrast, water vapor in the cyclone's inner region was isotopically enriched.” Since Nepartak was moving systematically toward Taiwan, the shape of the $\delta^{18}$O variation tracked the inner structure. Overall, a decrease from $-13\%e$ to $-19\%e$ implies a hypothetical single-stage rainout of 49% from the original vapor mass (see later Section 4; Malkus & Riehl, 1960; Yang et al., 2011).

### 3.4. Phase Lag of the d-excess Values of Nepartak Vapor

There is a large phase lag in the d-excess profile relative to the $\delta^{18}$O profile in case of Nepartak (see inset of Figure 5). The $\delta^{18}$O decreases as the typhoon Nepartak approaches the sampling station but the d-excess retains its typical oceanic value of $\sim10\%e$. As the typhoon disintegrates over Taiwan, the $\delta^{18}$O returns back to the pre-typhoon values but the d-excess decreases (from about 10%e to 5%e). The time difference is of the order of 6 days. This can be explained by a significant contribution of ground vapor in the aftermath of typhoon disintegration. After the arrival of typhoon over Taiwan and the associated enormous rainfall during the landfall (from...
7th to 13th July), evaporation and transpiration from ground and vegetation could be an important contributor to the overhead vapor. During the time period of July 6–9, 80 rainwater samples were collected. The average values of δ¹⁸O and d-excess of rainwaters are, respectively, −5.1 ± 2.1‰ and 3.2 ± 3.9‰ (details not shown here; Figure 1). As shown by Uemura et al. (2008), the δ-excess in vapor correlates negatively with relative humidity. Post-landfall, the ground is wet and the land evaporation takes place in a highly humid environment which makes the δ-excess minimum of vapor close to the rainwater value. Eventually, the d-excess as well as the δ¹⁸O values return to the oceanic values when the air mass is replaced completely by the southeasterly monsoon flows that happens by about July 24 (inset of Figure 5).

3.5. Typhoons Meranti and Malakas

Meranti started as a tropical depression over the Pacific Ocean in the afternoon of September 10, 2016 and gradually moved northward Cambay. It reached the status of super typhoon on September 12 at 11 a.m., with maximum sustained winds of 62 m/s and a central pressure of 935 mbar. Based on the infrared images shown in Figure 4, the outer rainbands of typhoon Meranti hit Taiwan on September 13, at 07:00 hr. The formation of eye took place on September 14, 03:00 hr (MR2; Figure 4). As the typhoon progressed toward the northwest direction, the inner rainbands of the typhoon reached the southern part of Taiwan on September 14 at about 00:00 hr. Within about 2 hr of hitting the southern coast, the eye of Meranti collapsed. The study area continued to receive rainfall from the outer rainbands, throughout the traverse of Meranti over Taiwan. It subsequently entered mainland China on September 15 and gradually degraded into a tropical depression. Meranti is one of the most intense tropical cyclones on record in the Western North Pacific Basin (according to the Japan Meteorological Agency, JMA, it had 220 km/hr maximum wind speed and 890 mb as the minimum central pressure; Source: Western North Pacific Typhoon Best Track File 1951–2022).

As typhoon Meranti was developing, another tropical depression, Malakas was forming in its wake. In fact, Figure 1 shows that Meranti itself formed in the aftermath of an intense tropical storm (the most intense as seen in its δ¹⁸O signature being the lowest encountered). Malakas made its way to the northwest at the beginning and turned northward on the southeastern side of Taiwan on September 17 and then turned northeast on September 18. These two typhoons were close to each other and formed almost in the same region affecting similar areas. Meranti moved quickly over the ocean in contrast to Malakas which had a tortuous journey. The meteorological parameters measured at Taipei station during the period September 7–21 when Meranti and Malakas developed in the west Pacific and moved to Taiwan-China region are shown in Figure 6. The tracks of these two typhoons (Figure 2), along with the cumulative rainfall (obtained from ERA5 product), are shown in Figure S4 in Supporting Information S1. They show that the most intense rainfall occurred in the region near the land. Since Meranti and Malakas came within a few days of each other their vapor isotope signatures were intermingled and the two individual V-shaped profiles were present but not fully developed.

3.6. Typhoon Megi

Megi formed as a tropical disturbance over the western North Pacific Ocean on September 21, 2016 and gradually moved northward Cambay. A period of steady intensification followed as Megi approached Taiwan. On September 24, Megi was trying to form an eye but stopped intensifying further till September 25. With good radial outflow tapping into westerlies and a large tropical upper tropospheric trough cell to the east, a defined eye remained absent but it could be categorized as typhoon. Megi strengthened to reach sustained winds of 185 km/hr hour at 18:00 hr, September 26 with a central pressure of 940 mb. On September 26, Megi was undergoing an eyewall replacement cycle (MG2 and MG3; Figure 4). After completion of this cycle, which resulted in a ragged but larger eye, Megi was able to gain intensity while making landfall in Taiwan. With maximum sustained winds of 52 m/s and a central barometric pressure of 935 mbar, Megi made its landfall in Hualien County, Taiwan on September 27 at 04:40. This storm was one of the strongest typhoons to make landfall in Taiwan with isolated

Figure 5. The meteorological parameters measured at Taipei station during the period July 1–19 when Nepartak developed in the West Pacific and moved to Taiwan-China region. Note that the minimum pressure and maximum wind speed occurred within July 8 and 9, 2016 but the δ¹⁸O minimum occurred about 2 days later. This shows that the δ¹⁸O distribution inside a moving typhoon vortex is not uniform; the δ¹⁸O minimum is not coincident with the eye position but displaced toward the posterior part. The δ-excess minimum at 17 July is displaced in time (see inset) indicating a phase lag of about 6 days of the d-excess minimum relative to the δ¹⁸O minimum. The depletion in d-excess values at around 19th July is explained by significant contribution of evaporotranspiration from the wet land after the typhoon landfall and associated heavy rains.
Figure 6. The meteorological parameters measured at Taipei station during the period September 7–21 when two typhoons (Meranti and Malakas) successively developed in the West Pacific and moved to the Taiwan-China region. MR0–MR2 and ML1–ML3 denote the respective positions of eyes of (see Figure 4) these two typhoons with respect to the measurement station.
rainfall amounts of 1,300 mm reported in Yilan County and 945 mm reported in Taipingshan (Figure S5 in Supporting Information S1).

Subsequently, interaction with the high mountains of central range of Taiwan caused Megi to weaken significantly and then enter into the Taiwan Strait at 13:10 on September 27. The meteorological parameters along with the isotopic ratios and the d-excess are shown in Figure 7. The minimum pressure and maximum wind speed were nearly coincident with the δ18O minimum around September 27. This must be due to the coincident arrival of the vapor signal and wind stress signal that happens in the eye region, suggesting that the isotope minimum occurred near the central zone. Figure S5 in Supporting Information S1 shows cumulative rainfall along the track on September 28 when Megi was passing over Taiwan. It shows that most intense rainfall occurred over the land rather than on the open sea.

4. Discussions

4.1. Estimation of Precipitation Efficiency of Nepartak and Other Typhoons

We show below that the fraction of vapor mass converted to rain (precipitation efficiency, PE, or inverse of characteristic moisture turnover time) is the most important parameter in determining the typhoon isotope ratio. For presentation and discussion purposes, we define PE (in percentage) as the ratio of rainfall P (mm/hr) accumulated over t hours and the averaged total precipitable water vapor (TPWV in mm) over the same time period, that is, PE for 1 hr duration and PE for 6 hr. Inverse of PE is the turnover time which varies from 1.6 to 3 days for the four typhoons considered. A tropical cyclone is an efficient engine that converts liquid to moisture at the lower level and moisture to liquid in the updraft. Figure S6a in Supporting Information S1 shows the distribution (spatial pattern) of cumulative rainfall over the track of Nepartak from July 2 to 9, 2016. The cumulative rainfall has a patchy structure with high value when the cyclone makes landfall in the southeast coast of Taiwan. After landfall (July 8, 2016), there is an efficient uplift of moisture caused by the Central Mountain Range with an altitude as high as 3,860 m (Figure 8). The intense cooling during this updraft converts the vapor efficiently to condensates.

To give an idea of the precipitation efficiency PE, we estimated the total precipitable water vapor (TPWV in mm) and rainfall (P in mm/hr) of the typhoon Nepartak at 16 points (Figure S9 in Supporting Information S1) over an area of ∼100 km radius circle along its track (from ERA5 reanalysis data). Hourly rainfalls over five circular zones at five chosen locations (zones 0, 3, 7, 11, and 15) along the Nepartak track are shown as colored Gaussian-like curves in Figure S8 in Supporting Information S1, as the typhoon moves from its inception to disintegration. The increase of the rainfall to a peak value followed by a drop embodies the variation as the typhoon passes over a given circular zone. As mentioned, there is also a sharp increase in rainfall as it approaches the land due to a more efficient uplift and condensation (shown in Figure 8 and Figure S8 in Supporting Information S1). We described earlier that during the movement, the typhoon scavenges the moisture from the zonal atmosphere leaving it slightly drier; later on, it is replenished by oncoming fresh moisture from larger distances. This phenomenon was tracked by plotting the TPWV at a given circular zone as a function of time as it crosses the zone. Results for five successive zones (0, 2, 4, 7, and 8) are shown in the top panel of Figure 9a. The bottom panel (Figure 9b) shows the gradual depletion and recovery in zone 3, where the average TPWV value before typhoon ingress is 61 mm; it falls down to 51 mm when the typhoon arrives and goes back again to the earlier value in 1 day when it leaves the area. It is also seen that during the initial phase of the typhoon, the relative depletion is high which becomes progressively lower in subsequent zones as the typhoon moves toward the land. The reason for this change is moisture build-up in the front region due to forward venting of moisture by the spinning circular arms. This is also the reason for a gradual increase in the TPWV value as the typhoon approaches the land. Before the onset of the typhoon, all zones have the same TPWV but after the onset more and more liquid water is converted to moisture by the typhoon engine. The insets in Figure 9a show the circular zones (0–16) considered in the model calculations and the inset in Figure 9b shows a simple visualization of the precipitation efficiency (PE) or fractional conversion of TPWV to rainfall in a time interval of 6 hr. Figure S9 in Supporting Information S1 shows the TPWV and average rainfall P of Nepartak over each of the 16 zones with radii of 100 and 500 km when the typhoon eye reached that zone, based on the ERA5 data.

To gain further insight into the spatio-temporal variability of PE, the PE of all four typhoons are shown in Figure S10 in Supporting Information S1 for zone radius of ~100 km. The ratio increases as typhoons approach the
Figure 7. The meteorological parameters measured at Taipei station during September 26–29 when the typhoon Megi developed in the West Pacific and moved to the Taiwan-China region. MG1–MG6 denote the respective positions of eyes of this typhoon with respect to the measurement station.
The moisture content is taken from ERA5. Nepartak gave rise to huge precipitation due to the efficient uplift of moisture. The moisture content is taken from ERA5.

Vertical radar picture of moisture distribution with altitude for typhoon Nepartak. After landfall (July 8, 2016 UTC) near the east and south coast of Taiwan, Nepartak gave rise to huge precipitation due to the efficient uplift of moisture. The moisture content is taken from ERA5.

The δ18O zone generally occurs without fractionation and the subsequent evaporation of it produces an elevated δ18O value (Laskar et al., 1996). The isotope ratio of this vapor is typically of oceanic origin (δ18O value about −12‰) but may be slightly depleted (about −13 or −14‰) when there are rainouts on the way. The δ18O values of western Pacific Ocean surface vapor ranges from −11‰ to −13‰ (Figure S2 in Supporting Information S1). On land, addition to this vapor comes from land surface evaporation and transpiration that tends to lower the δ18O value because the land vapor is ultimately derived from precipitation, which has lower δ18O values than ocean water (Laskar et al., 2014; Rozanski et al., 1993). However, near the coast, as in case of Taipei, the spray in the surf zone generally occurs without fractionation and the subsequent evaporation of it produces an elevated δ18O value of vapor (Fudeyasu et al., 2008) which can be carried inwards by sea breeze. Now in case of a typhoon, there is an internal dynamic of moisture recycling (shown for an idealized typhoon in Figure S1 in Supporting Information S1). There are some regions where there is an ascending motion of the air mass and there are adjoining regions where there is descent. During the uplift, the air cools and the vapor condenses to liquid phase, which may be brought down as rain. These raindrops exchange with the vapor at lower levels and further deplete the incoming vapor. Overall, the standing crop of vapor inside gets depleted more and more with time. We also noted that the internal vapor storage of a typhoon continuously receives contribution from fresh evaporation. However, Malkus and Riehl (1960) showed that the oceanic source of water vapor, apparently so vital for the very existence of a storm, makes only a small contribution to its water budget (within the typhoon radius). They indicated that the evaporation from ocean contributes less than 10% of the total inflow. In contrast, the rainfall associated with a typhoon is several times larger. According to Lawrence and Gedzelman (1996), for long-lived storm systems such as tropical cyclones, the total mass of rain produced over its lifetime is much larger than the instantaneous mass of vapor in the storm.

Typhoons

Lawrence et al. (2004) indicated that extremely low δ18O values in vapor occur in or downwind from regions of organized mesoscale weather disturbances (like typhoons) and can be as low as −25‰, much lower than the equilibrium isotopic value of vapor with seawater (Sánchez-Murillo et al., 2020). The mean δ18O value of vapor over the sea surface therefore decreases as storm activity and organization increase. They also showed that the stable isotope ratios of rain (Lawrence & Gedzelman, 1996) and water vapor (Lawrence et al., 1998) are usually markedly lower in tropical cyclones than in isolated thunderstorms due to their higher precipitation efficiency compared to the storms.

As the typhoon moves along its track and rains out (shown for the typhoon Nepartak in Figure S8 in Supporting Information S1 for ∼100 km zone), the isotope ratios of the typhoon vapor mass decrease in a systematic fashion. This is reflected in the left limb of the quasi-symmetrical V-shape (see Figure S5 for Nepartak) in the time variation of the vapor isotope ratio. This decrease can be explained by the following mechanism. During the summer typhoon season (JJAS), the air mass carrying atmospheric moisture moves mostly from the south or the east (Figure S2 in Supporting Information S1). The isotope ratio of this vapor is typically of oceanic origin (δ18O value about −12‰) but may be slightly depleted (about −13 or −14‰) when there are rainouts on the way. The δ18O values of western Pacific Ocean surface vapor ranges from −11‰ to −13‰ (Figure 1). On land, addition to this vapor comes from land surface evaporation and transpiration that tends to lower the δ18O value because the land vapor is ultimately derived from precipitation, which has lower δ18O values than ocean water (Laskar et al., 2014; Rozanski et al., 1993). However, near the coast, as in case of Taipei, the spray in the surf zone generally occurs without fractionation and the subsequent evaporation of it produces an elevated δ18O value of vapor (Fudeyasu et al., 2008) which can be carried inwards by sea breeze. Now in case of a typhoon, there is an internal dynamic of moisture recycling (shown for an idealized typhoon in Figure S1 in Supporting Information S1). There are some regions where there is an ascending motion of the air mass and there are adjoining regions where there is descent. During the uplift, the air cools and the vapor condenses to liquid phase, which is associated with isotopic fractionation. The liquid phase is enriched and the vapor is depleted. The same vapor descends down in the adjoining regions and mixes with the inward moving air mass. In addition, the liquid phase may be brought down as rain. These raindrops exchange with the vapor at lower levels and further deplete the oncoming vapor. Overall, the standing crop of vapor inside gets depleted more and more with time. We also noted that the internal vapor storage of a typhoon continuously receives contribution from fresh evaporation. However, Malkus and Riehl (1960) showed that the oceanic source of water vapor, apparently so vital for the very existence of a storm, makes only a small contribution to its water budget (within the typhoon radius). They indicated that the evaporation from ocean contributes less than 10% of the total inflow. In contrast, the rainfall associated with a typhoon is several times larger. According to Lawrence and Gedzelman (1996), for long-lived storm systems such as tropical cyclones, the total mass of rain produced over its lifetime is much larger than the instantaneous mass of vapor in the storm.

What is the main source for this rainfall? The water budget analysis of typhoon Nari and Morakot (Huang et al., 2014; Yang et al., 2011) have shown that the net influx (inflow minus outflow) constitutes the major part of the net condensation (precipitation), about 85%–90%, and only 10%–15% (Vanniere et al., 2020) is contributed by fresh evaporation. This indicates that the horizontal vapor transport supplies nearly all of the net condensation. The outflow is about 5%–15% of the total inflow, with the latter being composed of recycled water vapor, evaporation, and the atmospheric vapor the typhoon scavenges in its track. The strong low-level radial inflow transports highly moist air parcels from the surrounding oceanic environment to the eye region. The recycling is done by the air mass descent inside the typhoon body as well as from the long arms of the spiral bands (Figure S1)
The outflow at the higher level can carry lofted hydrometeors with low isotope ratios into regions of stratiform precipitation far from the storm center (Lawrence & Gedzelman, 1996). According to Tao et al. (2017), about 55% of the total rainfall in the inner-core region of tropical cyclones is contributed by stratiform rain. In spite of a much lower rainfall frequency, the convective rain quantitatively has a similar total volumetric rain as stratiform rain due to its higher rain rates. As mentioned before, according to Lawrence and Gedzelman (1996), the inward decrease of isotope ratios of vapor in tropical cyclones is due to recycling of water. As the down-drafted air passes beneath each rain band, falling raindrops evaporate and undergo diffusive isotopic exchange with the inward moving vapor reducing its isotope ratio. For a series of rain bands in a tropical cyclone this process results in a sequential lowering of isotope ratios of the vapor.

The access to water vapor relies on the typhoon's motion; as it moves through the atmosphere, the typhoon entraps the water vapor it encounters. Using a quantitative analysis based on MERRA (Modern-Era Retrospective analysis for Research and Applications) and TRMM (Tropical Rainfall Measuring Mission) data, Makarieva et al. (2017) have shown that for the Atlantic region, hurricanes sustain their rainfall along the track (over the area of their footprints) by collecting water vapor from a large circular area whose radius could be as large as 700 km. Some of this water is recycled vapor with lower isotope ratios. In effect, the typhoon is an extremely efficient
moisture-to-precipitation converting engine, which renders the left-over moisture isotopically depleted depending on how long the engine operates.

During the 2016 typhoon season, the wind was mostly from the south in June (Nepartak) and from the east and southeast in September (Meranti, Malakas, and Megi; Figure S2 in Supporting Information S1). The winds at lower levels were passing through the vapor mass at the boundary layers and bringing the oceanic vapor to Taiwan. After the start of a typhoon (say Nepartak), the winds started to bring the typhoon influenced vapor with depleted ratios. The isotope ratios of the vapor at the Taipei station thus decreased slowly as the typhoon was getting closer. This happened due to two factors: the typhoon itself was maturing with progressive rainout and vapor recycling (two contributors of the “age effect” a la Gedzelman et al., 2003) and the wind was scavenging more of the proximal vapor as it was coming closer to Taipei. As a result, the period of closest approach was marked by a minimum in the $\delta^{18}O$ values. When the typhoon moves away, the wind can only scavenge the left-over vapor in its wake whose contribution decreases with time as it gets diluted by the oncoming fresh oceanic vapor. This causes the slow rise. It is obvious that a symmetric V-shape would arise only when the center of the typhoon passes over our measuring station (like Megi or Malakas). If the typhoon passes far away (like Nepartak and Meranti), the symmetry is distorted and an asymmetric V-shape is developed. This is because when the typhoon disintegrates, the isotope memory in the local environment is gradually erased; how fast this erasure takes place is determined by the rate of incoming fresh oceanic vapor and its mixing with the old vapor in the environment near the station. In case of Megi, the drop and rise happened in 2 days for both sides (symmetric) but for Nepartak, the drop occurred from July 7 to 11 (4 days), whereas the rise happened from July 11 to 19 (8 days). This is also evident from a comparison of the closest distances of the eyes of the typhoons from the measuring stations (Table S2 and Figure S11 in Supporting Information S1). The closest distance between the eyes and the station for Malakas and Megi was rather small, 150 and 130 km, respectively, whose timings nearly coincided with those of the $\delta^{18}O$ minimum. Both show near-symmetrical isotope profiles. On the other hand, the $\delta^{18}O$ minimum for Nepartak occurs nearly 63 hr after the eye passed away from the closest spot (330 km) producing a strongly asymmetric isotope profile. Surprisingly, the $\delta^{18}O$ minimum in Meranti occurs nearly 40 hr before (predates) the eye reaches closest to the station (400 km). In this context, it must be noted that Meranti developed immediately after a very strong tropical depression that produced the lowest $\delta^{18}O$ value ($-24.7‰$) in the entire year of 2016 (Figure 1). The cause of such depleted value in this depression must be due to intense rainout but cannot be quantified due to its widely dispersed physical state. It seems that Meranti arrived near the station when the ambient vapor was already low in isotope ratio.

The typhoon effect on vapor isotope ratios at a given place (like Taipei) also depends crucially on whether the wind system reaching the station is sampling the typhoon-influenced vapor from regions at lower levels. If the typhoon is from east and the wind is from south it may not sample the vapor at all and we may not see the reflection of its presence in the vapor isotope variation. Therefore, it is difficult to predict how representative the isotope ratio variation of the typhoon strength or its recycle efficiency. We therefore discuss only those typhoons here where the air mass could sample the typhoon vapor properly.

4.3. A Simple Model to Explain the Isotope Variation: The Basic Idea

It is known that circulation of low-level frictionally driven inflow, upward vertical motion in central areas forming the clouds, and outflow at upper tropospheric levels constitute the basic air movement in a typhoon. As mentioned, Lawrence et al. (2002) have shown that water vapors from storms of longer duration and better organization have lower isotope ratios due to the continuous addition of fractionation effects associated with the rainout. This is possible because cyclones can vent large quantities of vapor to their surroundings at the upper levels that carry the memory of previous episodes of rainout depletion. They also showed that the isotopic ratios of the vapor at the ground level decrease with time in a particular typhoon. Three major factors determine the time variation of the vapor isotope ratio measured at a station away from the typhoon: (a) how much time it takes to reach the station (an age effect involving rainout), (b) how far from the station the typhoon is located at a given time, and (c) what is the delay in transporting the vapor signature under the given wind speed and direction. After the disintegration of the typhoon, the isotope ratio returns to its previous value by mixing with the fresh oceanic vapor. A proper explanation of the isotopic features of vapor at a given station arising out of a particular typhoon moving in its track requires a circulation model of water vapor with several isotopic processes (Gedzelman & Arnold, 1994) that is beyond the scope of the present work. However, a simple scenario can be constructed to understand the
basic mechanism based on Rayleigh fractionation during condensation using the rainfall over the track and the TPWV causing the decrease in the isotope ratio (left limb of the V-shape) and mixing of fresh oncoming vapor with the left-over moisture causing the increase (right limb). Such calculations are possible only for an idealized typhoon in terms of its dynamics. Nepartak typhoon is close to such an idealized case and we use the data for Nepartak to introduce this calculation. We then apply the same model to the other three typhoons with various simplifying assumptions.

4.3.1. Mass Conservation

Cyclones are visible in satellite pictures (Figure 4 and Figure S1 in Supporting Information S1) as cloud-covered circular regions with radii ranging from 100 to 1,000 km. Their heights generally vary from 15 to 20 km (Emanuel, 2003). We imagine an idealized cyclone as a cylindrical zone of spinning vortex moving over the ocean from its formation point to the disintegration region maintaining its shape and moisture content nearly constant (Figure S12 in Supporting Information S1). The speed is generally 20–25 km/hr. It sheds enormous amount of rain (typically 1%–2% of the total moisture content of the vortex in 1 hr over 500 km zone) while on its track. We assume that it replenishes the lost moisture by trapping atmospheric water vapor it encounters around the track through its cross-sectional area (Makarieva et al., 2017). In addition, water evaporating from the oceanic footprint region is also mixed. We assume that the left-over water vapor (~99%) is depleted during the rain-out through Rayleigh condensation while the fresh atmospheric vapor and the evaporated moisture (together ~1%) have composition close to the initial value. Air spirals into the storm at low levels (Figure S1 in Supporting Information S1), with much of the inflow confined to a shallow boundary layer, typically 500 m to 1 km deep. These three components sustain the total moisture content of the cyclone.

As discussed before, we obtained the TPWV (in mm) and rainfall (P in mm/hr) from ERA5 for Nepartak over 16 zones (Figures S7 and S8 in Supporting Information S1) along its track and found that within a circular region of 100 km radius the hourly accumulated P is about 8% of the TPWV (Table S2 in Supporting Information S1). This extraction percentage depends on the area of the region considered. Without any replenishment and recycling, the cyclone would lose its total mass in a matter of 10–20 hr if we consider the loss rate over 100 km area. However, the track over the warm ocean allows sufficient inflow of moist oceanic air masses to sustain the mass and energy requirement of the cyclone. The assumption of vortex conservation is supported by the intact circular structure of Nepartak (Figure S12 in Supporting Information S1) over most of the track as revealed by radar reflectivity maps with distinct rain bands. The structure breaks down only when it approaches the land mass of Taiwan and China.

4.3.2. Region of Moisture Import

We note that both rainfall and TPWV change as a function of distance from the track of the center but at different rates. Therefore, the average values of these two quantities change as we increase the circular zone radius within which the average is taken (Table S2 in Supporting Information S1). If we increase the radius of the circular area to 250 km, the fraction (PE) reduces to 2%–3% because the integral rainfall decreases steeply as we move away from the track but the TPWV does not change proportionately. For example, over 100 km radius, the average TPWV is 73.2 mm while for 250 km radius the value is 60.0 mm. Since P/TPWV varies with the choice of radius, we have to decide an appropriate size for the model calculation. Additionally, the replenishment rate (defined as evaporation plus advection) also varies with radius (Table S2 in Supporting Information S1). Makarieva et al. (2017) have shown that 700 km radius is a good choice for the moisture budget in Atlantic cyclones because water vapor available to the hurricane from within ~700 km from the hurricane center can make it self-sustaining.

Based on observations of the northwest Pacific cyclones, Frank (1977) showed that at a distance of 440–660 km from the cyclone center there is a region of air subsidence throughout most of the troposphere. Consequently, a distance greater than about 500 km corresponds to the so-called “moat of clear skies” that typically surround tropical cyclones. The zone within this 500 km radius contributes the water vapor drawn into the cyclone that descends to the lower levels to be driven inwards (Figure S1 in Supporting Information S1). This in-drawn moisture then condenses when lifted with the updraft to form the rainfall. We also note that the meteorological picture of the Nepartak (Figure S12 in Supporting Information S1) supports that a region within about 500 km from the center would be appropriate because beyond this distance, the condensed vapor becomes significantly less dense. We therefore assume that the rainfall is predominantly sustained from the atmospheric stock within 500 km of the cyclone center. Additionally, this size does not change and the typhoon has an organized, closed circulation that recycles and progressively fractionates the vapor. Since the rainfall/water vapor ratio decreases
with distance from the cyclone center, the vapor isotope ratios at the ground level would increase as we move away from the central zone. However, for simplicity, we assume that the whole moisture stock inside the cyclone has uniform δ^{18}O value. This would be possible due to the rapid mixing of the cyclone vapor by high-speed vortex whereby any internal variation is quickly averaged out. The choice of 500 km radius for calculation will be further discussed after the model performance is evaluated.

4.3.3. The Left-Limb of the V-Shaped Variation: Fractionation Calculation

Isotopic depletion in a vapor mass due to Rayleigh condensation depends on the fraction that rains out. In a system that is continuously raining and moving, the instantaneous isotope mass balance can be expressed in a differential form (Criss, 1999). We use the ratio $R_i$ and $\delta_i$ for typhoon vapor, $R_e$ and $\delta_e$ for rain, and $R_{in}$ and $\delta_{in}$ for all external vapor combined input, including atmospheric import and evaporation from the ocean. Considering $M$ (mm) as the total mass of the typhoon vapor, which is nearly equal to the mass of the $^{16}$O vapor, $P$ as the precipitation (mm/hr) inside the typhoon, and $X$ as the rate of the total imported vapor mass (mm/hr), we can write

$$M \frac{dR_i}{dt} = XR_{in} - PR_i \quad (1a)$$

$$M \frac{d\delta_i}{dt} = X\delta_{in} - P\delta_i \quad (1b)$$

To write Equation 1b from Equation 1a, we assumed $X \sim P$ since $M$ is nearly constant (62 ± 2 mm). Using the definition of isotopic fractionation factor ($\alpha$ is taken to be 1.01057 as discussed later)

$$\frac{R_i}{R_e} = \alpha \quad (2a)$$

$$\delta_i = a\delta_e + (\alpha - 1) \quad (2b)$$

Equation 1b can be re-written as

$$M \frac{d\delta_i}{dt} = X\delta_{in} - P[a\delta_e + (\alpha - 1)] \quad (3)$$

Using Equation 3, we can derive the $\delta_i$ values numerically in steps of 1 hr ($\delta t = 1$ hr) by plugging in the values of $M$, $P$, and $X$ as well as estimated value of $\delta_{in}$ obtained through assumptions described below. The hourly values of $M$ and $P$ at typhoon positions are obtained from ERA5.

4.3.3.1. The Value of $\delta_{in}$: The Impact of Evaporation

As discussed before, the total vapor import to the typhoon comprises of two components: evaporation (flux $F_e$) and atmospheric convergence (flux $F_c$) and corresponding delta values $\delta_e$ and $\delta_c$. Therefore, $\delta_{in}$ is the sum of two constituents (evaporation and convergence flux)

$$\delta_{in} = k_e\delta_e + (1 - k_e)\delta_c \quad (4)$$

where $k_e = F_e/F$. Based on the discussion in Section 4.2, we take 0.2 as an estimated value of $k_e$ applicable to all four typhoons (A. Makarieva, personal communications). The value of $\delta_e$ (the delta value of the evaporating flux) has not been measured for the Pacific Ocean (especially for typhoons) but since its contribution is small, we can use the value for the Atlantic Ocean measured by Benetti et al. (2017). Based on an extensive analysis and modeling, they give the mean $\delta_e$ as −6‰. We also take $\delta_c = \delta_a$, where $\delta_a$ is the atmospheric delta value (about −12.3‰ for Nepartak). There are some minor differences among the typhoons as given in the tables showing the calculations given in the Supporting Information. Using these values, $\delta_{in}$ turns to be about −11.0‰, close to but slightly different from $\delta_a$ (−12.3‰).

We see that it is essential to know $k_e$, $\delta_e$, and $\delta_c$ to calculate $d\delta_i/dt$. An order of magnitude estimate for the total decrease in the delta value would be instructive to show this. For Nepartak, $\delta_c$ decreases by $\Delta\delta_c = -7‰$, from −12.3‰ to −19.3‰, in $\Delta t = 96$ hr. Because $M$ does not change significantly (62 ± 2 mm; Table S3 in Supporting Information S1), we take input flux = rainfall, that is, $X = P$ in mm/hr in Equation 1 and mean $P/M = 0.014$ hr⁻¹ (from Table S3 in Supporting Information S1). Using the mean values of $\delta_c = -16‰$ in
Equation 3 with $\delta_{in} = -11\%$ and $(\alpha - 1) = 10.6\%$, we obtain $d\delta_i/dt = (-5.4\%) \times 0.014$ and $\Delta \delta_i = (-5.4\%) \times 0.014 \times \Delta t = -7.3\%$, which is close to the observed value.

Using $X = P$ and recasting the equation in term of $MIP = \tau$ in unit of hr$^{-1}$

$$\tau \frac{d\delta_i}{dt} = -\alpha \delta_i - (\alpha - 1) + \delta_{in}$$

This equation reveals that there is a steady state value beyond which $\delta_i$ would not decline further. Indeed, as the typhoon becomes depleted in $^{18}O$, constant rainfall removes less and less $^{18}O$ per unit time. At a certain point, $^{18}O$ depletion will be balanced by $^{18}O$ import. The steady state value can be obtained by equating the left-hand side to be zero and solving for $\delta_i$: $\delta_i = -1.0106 \times \delta_i - 10.57 - 11.0$, and so, $\delta_i$ (steady state) = $-21.57/1.0106 = -21.3\%$. This value is quite close to the minimal value of $-19.3\%$ obtained for Nepartak. It seems that a typhoon like Nepartak approaches a steady state isotope balance within about 4–5 days after its inception by the massive rain out and equivalent import.

4.3.3.2. Impact of Air Exchange

For the sake of completeness, there is one more caveat. Typhoon exchanges with the environment not only water vapor but also dry air. There is an air inflow (predominantly in the lower troposphere) and outflow (predominantly in the upper troposphere). Both inflow and outflow carry water vapor, which in the general case have different $\delta_v$. The inflow has $\delta_v$, the atmospheric value. The outflow has $\delta_v$, the typhoon value. The mean $\delta_v$ of water vapor convergence depends on the relative rates of water vapor gross import/export by the inflow/outflow:

$$\delta_v = \frac{\delta_v F_{in} - \delta_v F_{out}}{F_{in} - F_{out}} = k_{in} \delta_a + (1 - k_{in}) \delta_v$$

where $F_{in}$ and $F_{out}$ are the gross fluxes of inflow and outflow of water vapor to and from the typhoon, $F_{in} - F_{out} = (1 - k_v)P$ is water vapor convergence, $k_{in} \equiv F_{in}/(1 - k_v)P$.

The value of $k_{in}$ characterizes how resistant the cyclone is with respect to environmental mixing. If $k_{in} \gg 1$, strong mixing will prevent $\delta_v$ from deviating appreciably from $\delta_v$. Indeed, combining Equations 4–6 and re-arranging the terms we find

$$\tau \frac{d\delta_v}{dt} = -\delta_a (\alpha + k_{in} - 1) - (\alpha - 1) + k_v \delta_v + (1 - k_v) k_{in} \delta_a$$

From this, it is clear that with $k_{in} \gg 1$, the equilibrium minimal value is $\delta_{v_{eq}} = \delta_v$ that is, no decline of $\delta_v$. For the assumption $\delta_v = \delta_v_{eq}$ to work, the outflow must export a negligible amount of water vapor, such that $k_{out} \approx 1$ and $k_{in} - 1 \ll 1$ in Equation 7. This means that nearly all water vapor that is imported with the inflow, precipitates within the typhoon. It is not an implausible assumption, because water vapor concentration decreases sharply with altitude and is negligible where the outflow predominantly occurs. After all, the very fact that $\delta_{v_{eq}}$ does decline, testifies that $k_{out} \approx 1$. On the other hand, if the typhoon did not export its own water vapor at all, the question arises how then typhoon’s $\delta_v$ could be registered at a station outside the typhoon. Therefore, some export must occur, but in quantities not perturbing $k_{in} - 1 \ll 1$. Unfortunately, we cannot get an idea about the export from the typhoon vapor mass because the mass $M$ generally increases with its progress as seen from the numerical values (Figure 9).

4.3.3.3. Mean Temperature of Major Rain Formation Zone: The Value of $\alpha$

Fractionation factor, $\alpha$, depends on the temperature of rain formation region. We know that in a typhoon, it normally rains hard at the surface of the eyewall and spiral arm-bands. The low-level air convergence is mainly in the lowest 1 km but it rains all the way to the tropopause at over 10 km altitude (Houze, 2010). However, the main rains are in the lowest 2–4 km altitude (Ming-Jen Yang and Kevin Trenberth, personal communications) because the cloud layers are the thickest in this range (Palmen & Newton, 1969). Therefore, we should use the temperature at these altitude levels. There are several studies dealing with temperature profiles in typhoons (Lasota et al., 2020; Makarieva et al., 2017; Munsell et al., 2018) and comparison of those profiles with the west North Pacific SST during typhoon formation suggests a temperature range of 12 ± 2°C in the 2–4 km height range. To find the lapse rate, we also analyzed the ERA5 temperature variations along the Nepartak track at 800 and 1,000 mb (see the Discussion and Figure S6b in Supporting Information S1), which shows that within the 610–800 mb
levels (corresponding to 2–4 km altitude range), the average temperature would be about 285K (or 12°C). We, therefore, use this value as the mean condensation temperature to obtain the equilibrium alpha value of 1.01057 (Horita & Wesolowski, 1994) for our calculation.

4.3.3.4. Choice of Radius

The estimated value of the size or radius \( r \) of the typhoon is an important issue. Unlike the basic mass balance considerations above, it is specific to tropical storms. There may be a concern about why we had to assume a certain radius (i.e., 500 km) for which all the calculations were made. Our first response is that \( P/M \) changes very abruptly with radius at smaller radii, but then changes much less at larger radii, so a large radius is a good choice. However, this is not the whole picture. We note that \( d\delta_v/dr \) is proportional to \( P/M \). For a smaller radius, \( P/M \) is much higher, and from Equation 1, a significantly higher \( d\delta_v/dr \) would result that would not have fit the data. For example, for a radius of \( \sim 100 \) km, \( P/M \) is six times higher than for a radius of 500 km (Table S2 in Supporting Information S1). If, on the other hand, a larger radius was chosen, where \( P/M \) is somewhat smaller (\( P/M \) at a radius of 1,000 km is 20% lower than at 500 km), we would have obtained a lower \( d\delta_v/dr \). In addition to the impact of lower \( P/M \), there is additional effect due to increase of the role of evaporation (and decreasing role of moisture convergence), that is, an increase of \( k_v \) in Equation 7. According to Figure 10 of Makarieva et al. (2017), \( k_v \) rises from about 0.2 at 500 km radius to 0.5 at 1,000 km radius. By changing the radius where Equation 7 is evaluated between 100 and 1,000 km, it is possible to obtain \( d\delta_v/dr \) values that differ by more than 1 order of magnitude. Therefore, the question arises as to what are the physical grounds that make \( r \approx 500 \) km the radius of choice. One would guess that this radius should be close to the inner radius \( r_o \) of the storms “dry footprint”—the outer region where storm’s precipitation turns out to be smaller than climatological precipitation (e.g., Makarieva et al., 2017; Figures 4g and 7). This suppression of precipitation indicates predominantly descending air motion. For North Atlantic hurricanes, it is about 600 km.

Since the parameters of Equation 7 have considerable uncertainty, given the high sensitivity of \( d\delta_v/dr \) to \( r \), especially at smaller \( r \), one cannot expect to obtain a close match to observed \( d\delta_v/dr \) using the observed cyclone size reflecting the cloud cover. Rather, our choice is validated by the fact that the radius where the model shows the best fit is indeed fairly close to the characteristic radius where precipitation is reduced to its climatological values.

4.3.3.5. Numerical Calculations and Plots

We computed the \( P \) and TPWV ratios for the 96 hourly positions of the Nepartak cyclone but showed representative values for positions separated by 6 hr from the previous position (Figure S7 in Supporting Information S1). The 6-hr period is chosen because the total travel time of Nepartak over the ocean was 4 days (as estimated from the typhoon eye movement; 4 days = 96 hr = 6 × 16 hr; Figure S7 in Supporting Information S1) and the isotope variation plot consisting of 16 points gives a sufficiently smooth curve to compare with the actual measured data. For clarity, the details of our step-wise model are shown as a schematic in Figure 10 in 6-hourly steps. In the first 6 hr time step (denoted by index #0), the starting mass \( M_0 \) of the moisture rains out a fraction \( P_f/M_0 \) (about 9% in 6 hr leaving behind a vapor fraction \( \approx 0.91 \)); in the second step, the fraction \( P_f/M_1 \) is about 8% in 6 hr leaving behind a vapor fraction of 0.92 and so on (see Table S3 in Supporting Information S1).

Since the model values are calculated inside the typhoon whereas the measurement is done in Taipei, we must put in a delay to account for the isotope signal to propagate from its place of origin (the changing locations of the typhoon) and reach Taipei. From July 6 to 10, the wind direction was in general from east to west, thus facilitating the efficient transfer of the vapor isotope signal to Taipei (Figure S12 in Supporting Information S1). It is, however, difficult to decide how much would be a good choice as it would depend on the speed and direction of the air mass passing through the typhoon (which itself is moving) and carrying its signature to the Taipei station. We chose 30 hr as an average delay for Nepartak, which matches the general profile of the model values with the observation. The model was run from July 6, 2016, 2 p.m. to July 10, 2016, 8 a.m. and the predicted time variation (Figure 11a) shows a good match with the observed values as a function of time.

Megi was a short-lived typhoon where the left-limb developed in 36 hr. In addition, it was not an idealized typhoon in its signature. Satellite cloud picture shows that Megi has a radius of influence similar to that of Nepartak (Figure S13 in Supporting Information S1). Nepartak vapor signal was clear from July 5 to 19, 2016, whereas the signal for Megi was visible from September 26 to 29, 2016. The shorter duration is because Megi traveled much faster over its track compared to Nepartak (implying a less age effect). We used 3-hr time steps.
to obtain at least 12 model points for the plot (Table S5 in Supporting Information S1). The fit was done using −12.4‰ as the initial vapor isotope value and δin = −11.1‰ and a variable delay from 16 to 4 hr because the arrival times of the Megi signal were not uniform. Table S5 in Supporting Information S1 shows the model calculation values and Figure 11b shows the plot. The data points for the left-limb of Megi were not smoothly distributed but the model values make a reasonable fit.

The case of Meranti-Malakas is the most complex and difficult to model. The idealized model discussed above for the two well-developed typhoons Nepartak and Megi cannot be applied to the typhoons Meranti (September 12–16) and Malakas (September 16–20) in a straightforward manner. These typhoons not only occurred very quickly one after the other but also, to start with, they were riding on a huge storm system. The typhoon Meranti collected the vapor left from one unidentified previous storm (see Figure 1) with highly depleted isotope ratio (about −25‰) and was immediately followed by the typhoon Malakas before full disintegration. We took −13.3‰ as a hypothetical initial value for the atmospheric vapor and the δin = −11.8‰. Calculations were done with 6-hr time steps. Since the starting composition of Meranti was already depleted, the plot fits only the second half of the left-limb profile (Figure 11c). For Malakas again, the initial value was a virtual −13.3‰ with δin = −11.8‰. However, the Malakas measured values did not follow a smooth decrease. Considering the scatter, the 6-hr step time plot gives an acceptable fit to the data (Figure 11c). From these two cases, it is clear that the simplified method has limitation when applied to complex isotope profiles. We think the mismatch could be due to several factors. (a) The wind system for these two cases was not bringing in the oceanic vapors systematically to the measuring station at Taipei. (b) Malakas came close to Taiwan but swerved to right while at a distance of about 140 km, proceeding to Japan in the mid-ocean (Figure 2). Meranti also did not reach Taiwan but came at the closest distance of 400 km of Taipei while at sea. (c) The wind trajectories were also such that the vapor mass did not arrive at the monitoring station systematically.

The lesson that we learn from this simple model (albeit with a number of assumptions) is that the isotope evolution of the typhoon vapor mass is dominantly controlled by fractionation associated with Rayleigh condensation. The rainout causes continuous fractionation along the track, which varies with fraction rained out (from about 7% to 10% and on average about 8.5%) considering a 6-hr period). The rate of decrease in delta value depends on the factor P/M, the average hourly value (in hr⁻¹) of which varies from 0.014 to 0.023 (mean 0.017 hr⁻¹) for the four cases considered. This factor determines the slope of the left-limb. Megi with the slope of 0.023 hr⁻¹ has the steepest decline, a drop of −5.2‰ in 36 hr compared to a drop of −7.0‰ in 96 hr for Nepartak with a slope of 0.014 hr⁻¹. The typhoon vapor preserves the signal of depletion, which can add up to a large net fractionation (by

Figure 10. Schematic picture of the Rayleigh fractionation model to show how moisture recycling (mixing of the leftover typhoon vapor stock and imported fresh oceanic vapor flowing toward the center at low levels) leads to continuous depletion of heavy isotopes in the vapor of an idealized typhoon; percentage (%) denotes rain out amount in 6 hr relative to the typhoon vapor mass. The numbers are based on the data of the first two steps for the Nepartak typhoon as shown in Table S3 in Supporting Information S1.
Figure 11. (a) Variation of isotope ratio ($\delta^{18}$O) of the Nepartak vapor measured in Taipei along with the model prediction. The sharp decrease in the ratio over 4 days (from July 6 1400 hr to July 10 0800 hr) is reasonably reproduced by the Rayleigh condensation model (Figure 10) after putting in a delay of 30 hr for signal transfer. The subsequent recovery is also well reproduced by a simple moisture-mixing model. (b) Comparison of model predicted $\delta^{18}$O values of the typhoon Megi vapor with that observed. Note that we had to put in an average delay of 10 hr to the model times to account for the signal transfer. (c) Comparison of model predicted $\delta^{18}$O values of the typhoon Meranti and Malakas vapor with that observed. The details of the assumptions behind the calculations are given in the main text.
as much as $-7\%$ in case of Nepartak an idealized case) because the recycling of old vapor has more impact with only small changes caused by addition of fresh oceanic vapor.

The drop in $\delta^{18}O$ value for Nepartak occurred from July 5 to 9 (4 days), whereas for Megi it was from September 26 to 27 (less than 2 days). We find that the rainout for the two typhoons were similar (average 8.4%) but since Megi was active for a smaller period the depletion was smaller (from a starting composition of $-12.4$ to $-18.5\%$) and took place in a short time of about 36 hr. As in the case of Nepartak, the depletion signal could be carried to the sampling station due to the favorable wind direction from east scavenging the typhoon vapor to Taipei (Figure S13 in Supporting Information S1).

It is true that we have proposed a rather simplistic model without considering the complex interplay of up-draft, rainout, down-draft, and raindrop evaporation at various levels of a typhoon (Figure S1 in Supporting Information S1). However, the typhoon internal dynamics of fractionation by a simple multiple stage Rayleigh model seems sufficient to provide explanation for the isotope variation and shows the importance of precipitation efficiency.

### 4.3.4. The Right Limb of the V-Shaped Variation for the Four Typhoons

After the disintegration or departure of a cyclone from the neighborhood of Taiwan, the left-over vapor mass with low isotope ratios (about $-20\%$) slowly gets replaced by the oncoming oceanic vapor with higher isotope ratios (about $-12\%$). However, Taipei being located in a basin-like valley (Taipei basin; Figure S14 in Supporting Information S1), the mixing in the boundary layer is slow. To calculate the rate of replacement of the old vapor with fresh vapor, we construct a box around Taiwan having latitude and longitude ranges: $22^\circ$–$26^\circN, 119^\circ$–$123^\circE$ (Figure S15 in Supporting Information S1) where the vapor is partly mixed on a daily scale. In our case, the wind direction of the oncoming vapor was mostly from the south-west after a typhoon disappeared whereas they were mostly from the east when the typhoon was coming toward Taiwan. We can model the replacement by a simple mixing calculation with the following:

Let us assume that the ratio in the box at the initial stage is $\delta_{in}$, the final ratio is $\delta_{fin}$, $\delta_a$ is the incoming (including evapotranspiration) ratio and the mixing fraction is ($M$ is the vapor mass in the box):

$$\delta_{fin} = (1-x)\delta_{in} + x\delta_a$$

If the incoming flux is $F$, considering a time interval $\Delta t$, $x = F\Delta t$. This results in $\Delta \delta_{fin} / \Delta t = (\delta_a - \delta_{in})F$.

Casting the above as a time evolution differential equation, where $\delta_b$ ($\delta_b = \delta_a$) is the final value of the box, replacing $\delta_{in}$ varies with time:

$$d(\delta_a - \delta_b)/dt = -F(\delta_a - \delta_b)$$

whose solution is

$$(\delta_a - \delta_b) = (\delta_a - \delta_{b,\text{start}})\exp(-Ft)$$

As $t$ increases the $\delta_a - \delta_b$ $\to 0$ or $\delta_b \to \delta_a$ the $\delta^{18}O$ value changing from the starting box value ($\delta_{b,\text{start}}$) to the oceanic value at the end. The time evolution depends on the value of $F$, the flux of the incoming vapor. We estimate the value $F$ using the column-integrated vapor across the boundaries of the box and solve for $\delta_b$ using a numerical approach.

In case of Nepartak, the recovery date ranges from July 10, 2016 to July 19, 2016 and the average TPWV over box ($M$) during this period was $1.05 \times 10^{13}$ kg. We obtained average inward flux of vapor during 24 hr period as $8.9 \times 10^{12}$ kg per day through the south and west boundaries of the box (subtracting the rainout fluxes of $9.6 \times 10^{11}$ kg per day and evaporative fluxes of $6.1 \times 10^{11}$ kg per day). Almost similar amount was exiting through the east and north boundaries keeping a nearly balanced TPWV. The average incoming flux was thus about 85% of the average TPWV. However, not all of it mixes uniformly. Since the isotope data represent daily averages and the flux varies greatly during the day or even from 1 day to another, we simplify the calculation by considering averages of TPWV and the incoming flux for mixing calculation.

The composition of the incoming vapor is not known a-priori except that the pure oceanic composition is $-12.0\%$ from both our data and the global ocean vapor map of Good et al. (2015). As a guide to estimate the isotopic changes by the incoming vapor contribution, we note that in 6 days the $\delta^{18}O$ of the box vapor changes from a
value of $-19.3\%$e to $-14.1\%$e (about $5.2\%$e increase). Therefore, the average increase per day is $0.8\%$e (during these 6 days) which is about 15% of the total change of $5.2\%$e. By a trial and error method, we see that the oceanic component with $\delta^{18}O$ value of $-12.0\%$e should be about 15% of the incoming flux during the initial part of the recovery phase on a per day basis. As we mix, the box value increases each day, and the rate of increase slows down. The flux percentage has to be increased to compensate for this. Carrying on the calculation every 24 hr, and fixing the percentage of the incoming vapor flux in an increasing fashion (from 15% to 100%) by systematic trials, we could reproduce the right limb of the isotope variation reasonably well (Table S5 in Supporting Information S1 and Figure 11).

A similar set of calculations was done for Megi. In this case, the average TPWV value was $1.19 \times 10^{13}$ kg (slightly more than the Nepartak case) but the average Inward flux was higher, about $1.15 \times 10^{13}$ kg/day. Moreover, Megi had only $-17.7\%$e at its lowest point. Megi vapor distribution around Taiwan was not as extensive as Nepartak and the wind speed was much higher (about 20–25 m/s compared to 12–15 m/s for Nepartak). These factors resulted in a much faster recovery (about 2 days compared to 10 days for Nepartak). By an argument similar to the Nepartak case, we noted that by increasing the percentage component from 14% to 26% every 6 hours, we could increase the box isotope ratio from $-17.5\%$e to $-13.1\%$e and reproduce the right limb for Megi reasonably well (see Table S6 in Supporting Information S1 and Figure 11b).

Application of the mixing model for the right limbs in the Meranti-Malakas cases involved much trial and error because the rises are not smooth. Meranti rise is disturbed by the entry of Malakas. Our equations can only be applied in smoothly rising cases. The results are shown in Figure 11c (Tables S7–S10 in Supporting Information S1), which show that the disagreements are large, as expected in case of a dual system of typhoons riding on vapor left behind by an extremely strong storm system and mutually interfering.

5. Conclusions

High-resolution oxygen isotope ratios ($\delta^D$ and $\delta^{18}O$) and d-excess values in atmospheric vapor have been measured using a Picarro analyzer in a ground station in Taipei to decipher the moisture sources of a set of four typhoons (Nepartak, Meranti, Malakas, and Megi) that successively invaded Taiwan in 2016. The isotope ratios display large variations during the time when the typhoon vapor reaches the measuring station. After the disintegration of the typhoon, the ratio reaches back to the normal oceanic value. Together these changes (decrease followed by increase) have a V-shape but the structure differs from one typhoon to another.

The moisture-to-precipitation conversion rate expressed by precipitation efficiency, PE (rainfall/TPWV; Malkus & Riehl, 1960; Yang et al., 2011) is an important parameter that determines the vapor isotope ratio variation inside the typhoon vortex during the phase changes (water to vapor and vapor to condensates) due to isotopic fractionation. The inverse of PE is also known as the turnover time of moisture which varies from 1.6 to 3 days (on average about 2 days) in the case of the four studied typhoons. In addition, the input moisture to sustain the typhoon comes from the environmental vapor mass and recycling of the vented vapor caused by downdraft and lower level flow. This is an interesting issue because the oceanic evaporation is only a small part of the total precipitation. Makarieva et al. (2017) showed by a semi-quantitative analysis that by virtue of its steady movement, a given typhoon system appropriates a sizable proportion of the atmospheric vapor it encounters over its cross-sectional area. This mass contributes to the existing vapor which itself is continuously recycling. Using this concept of mass balance and the PE factor in a step-wise fashion, an evolution of the vapor isotope ratio inside the typhoon can be modeled as it moves from its origin toward the landfall area. The isotope ratios decrease as the typhoon evolves and the signal is transferred to the measuring station if there is a favorable wind.

By using European Centre for Medium-Range Weather Forecasts (ECMWF) fifth Re-Analysis (ERA5) derived Total Precipitable Water Vapor (TPWV), precipitation flux, and moisture recycling flux and adopting the Makarieva et al. (2017) scenario, we calculated the decrease in the isotope ratio ($^{16}O$/$^{18}O$) as a function of time in case of two major typhoons (Nepartak and Megi). Post-dissipation, the local vapor is replaced by the fresh oceanic vapor and the isotope ratio returns to its original value. The predicted V-shape agrees reasonably well with the observed signal in Taipei. The actual rainout takes place in multiple stages with about 9% rainout accumulated over a period of 6 hr (PE6) in each stage considering a 500 km radius circular area of rainfall. The close agreement of the model with observation in case of two well-developed typhoons (Nepartak and Megi) confirms that the buildup of isotopic fractionation by Rayleigh condensation can be tracked by using the precipitation
efficiency estimated from ERA5 data resources; in addition, a 500 km radius size seems quite appropriate for moisture budget of Pacific typhoons. Our study lends support to the observation that a continuous import of moisture by lower level inflow along the track matches the export through rainout and venting via upper level cyclonic arms.

Data Availability Statement

All the water vapor isotope data presented in this work are provided in the Supporting Information and also archived at Harvard Dataverse https://doi.org/10.7910/DVN/UTE3Q1 (2018). The meteorological data at Taipei station (site number 466920) are publicly available at Taiwan Central Weather Bureau. ERA5 hourly data are downloaded from the Copernicus Climate Change Service (C3S) Climate Data Store.

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