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Moist convection and water vapour transport into the upper troposphere over the Third Pole in multi-scale ICON-CLM simulations

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सार – तृतीय ध्रुव क्षेत्र (Third Pole) में ऊपरी क्षोभमंडल (Upper Troposphere–UT) के भीतर जलवाष्प के विनिमय की प्रक्रिया अभी भी पर्याप्त रूप से समझी नहीं गई है। पूर्ववर्ती अध्ययनों में मुख्यतः पुनर्विश्लेषण और उपग्रह आँकड़ों का उपयोग करके मानसून से संबंधित ऊपरी क्षोभमंडलीय आर्द्रीकरण पर ध्यान केंद्रित किया गया है। इस अध्ययन में तृतीय ध्रुव क्षेत्र में संवहन द्वारा जलवाष्प को ऊपरी क्षोभमंडल तक पहुँचाने की भूमिका की जाँच करने के लिए किलोमीटर-पैमाने ($D_x = 3.3$ km) के आईकॉन-सीएलएम (ICON-CLM) अनुकरणों का उपयोग किया गया है। विभिन्न वायुमंडलीय प्रक्रियाओं और ऋतुओं से संबद्ध तीन चरम वर्षण एवं हिमपात घटनाओं का विश्लेषण किया गया, ताकि मॉडल रिज़ॉल्यूशन और संवहन-योजना विन्यास का ऊपरी क्षोभमंडल में नमी परिवहन पर प्रभाव समझा जा सके। गहन संवहन पैरामीट्रीकरण (deep convection parameterisation) का उपयोग करने वाले बड़े ग्रिड वाले ($D_x = 13$ km) ICON-CLM अनुकरणों तथा पुनर्विश्लेषणों (IMDAA, ERA5) की तुलना में, किलोमीटर-पैमाने (km-scale) वाले ICON-CLM अनुकरण ने इन चरम घटनाओं (extreme events) का काफी बेहतर निरूपण प्रस्तुत किया। तिब्बती पठार (TP) के पूर्वी भाग में मध्य-पैमाने के तिब्बती पठार भंवर की घटना के किलोमीटर-पैमाने के अनुकरण में, पैरामीटरित उथले संवहन वाले प्रयोग में संवहन के स्पष्ट अनुकरण वाले, अर्थात् बिना किसी पैरामीट्रीकरण वाले, प्रयोग की तुलना में 300 hPa से अधिक ऊपरी क्षोभमंडलीय क्षेत्र में अधिक जलवाष्प परिवहन पाया गया। इसके विपरीत, मध्य हिमालय में मानसून के दौरान हुई चरम वर्षण घटना के लिए इसका उलटा परिणाम प्राप्त हुआ। सभी ICON-CLM अनुकरण तथा पुनर्विश्लेषण तिब्बती पठार पर झील-प्रभाव से प्रेरित चरम हिमपात की घटना को पकड़ने में विफल रहे और ऊपरी क्षोभमंडल में जलवाष्प में केवल मामूली वृद्धि प्रदर्शित की। सभी घटनाओं के लिए तिब्बती पठार के ऊपर किलोमीटर-पैमाने पर अनुकरण की गई ऊपरी क्षोभमंडलीय जलवाष्प उपग्रह प्रेक्षणों के अनुरूप रही। इसके अतिरिक्त, लैंग्रजियन वायु-पार्सल ट्रैकिंग और CALIPSO प्रेक्षणों से संकेत मिला कि हिमालयी क्षेत्र की चरम घटनाओं ने विश्लेषित अन्य घटनाओं की तुलना में ऊपरी क्षोभमंडल में अधिक जलवाष्प पहुँचाई। लैंग्रजियन ट्रैकिंग आरंभ होने के कुछ घंटों के भीतर केवल कुछ ही वायु-पार्सल क्षोभसीमा (100 hPa) के निकट तक पहुँच सके; ऊपरी क्षोभमंडल में पहुँचने के बाद ये वायु-पार्सल पूर्व की ओर प्रवाहित हुए। विश्लेषित तीन घटनाओं से संकेत मिलता है कि अनुकरणों में चरम घटनाएँ ऊपरी क्षोभमंडल में जलवाष्प के परिवहन को बढ़ा सकती हैं, किंतु तृतीय ध्रुव क्षेत्र के ऊपर संवहन के माध्यम से निचले समतापमंडल में प्रवेश आवश्यक रूप से नहीं होता। इसके अतिरिक्त, जाँच की गई घटनाओं के लिए, संवहन (convection) से जुड़ी ऊर्ध्वाधर परिवहन प्रक्रियाओं (vertical transport processes) को बड़े ग्रिड वाले, संवहन-पैरामीट्रीकृत सिमुलेशन और पुनर्विश्लेषणों की तुलना में किलोमीटर-पैमाने (km-scale) के अनुकरणों में अधिक सटीक रूप से निरूपित किया गया है।

ABSTRACT. Water vapour exchange into the upper troposphere (UT) remains poorly understood over the Third Pole region. Previous studies have mainly focused on monsoon related moistening of the UT using reanalysis and satellite data. In this study, we employ km-scale ($D_x = 3.3$ km) ICON-CLM simulations to investigate the role of convection in transporting water vapour into the UT over the Third Pole. We analyse three extreme precipitation and snowfall events, each associated with different atmospheric processes and seasons, to better understand the impact of model resolution and convection-scheme configuration on UT moisture transport. The km-scale ICON-CLM simulations provide a substantially improved representation of these extreme events compared to coarser ($D_x = 13$ km) ICON-CLM simulations and reanalyses (IMDAA, ERA5) that use deep convection parameterisation. A km-scale simulation with parameterised shallow convection of a mesoscale Tibetan Plateau (TP) vortex event over the east of TP showed more water vapour transport to the UT region (above 300 hPa) than a km-scale simulation without any parameterisation (i.e. with explicit simulation) of convection. The opposite was simulated for an event with extreme precipitation during monsoon over the central Himalayas. All ICON-CLM simulations and reanalyses failed to capture the extreme snow

event driven by lake effect processes over the Tibetan Plateau and exhibited only a slight water vapour enhancement in the UT. For all the events, km-scale simulated water vapour in the UT was consistent with satellite observations over the TP. Additionally, Lagrangian air parcel tracking and CALIPSO observations indicate that the extreme events over the Himalayan region transported more water vapour into the UT than the other analysed events. Only a few air parcels reached near the tropopause (100 hPa) within a few hours of the Lagrangian tracking initiation; once air parcels reach the upper troposphere, they advect eastward. The three analysed events suggest that, in the simulations, extreme events can enhance water vapour transport into the UT but may not convectively penetrate the lower stratosphere over the Third Pole region. Additionally, for the investigated events, the vertical transport processes associated with convection are better represented in the km-scale simulations than in the coarser, convection parameterised simulations and reanalyses.

Key words – Km-scale, Upper troposphere, Third pole, Water vapour, Deep convection.

1. Introduction

The upper troposphere and lower stratosphere (UTLS) have essential roles in Earth's atmosphere, especially Earth's energy balance (Uma *et al.*, 2014). Water vapour (WV), trace gases, and aerosols in the UTLS region can alter the incoming and outgoing radiation from the earth, which impacts regional and global weather and climate (Fadnavis *et al.*, 2013; Singh *et al.*, 2020, 2021). Increased water vapour leads to warming, affecting UTLS temperatures and dynamics (Charlesworth *et al.*, 2023). The transfer of moisture across the tropopause impacts stratospheric chemistry, particularly in ozone depletion processes. Aerosols in the UTLS region affect Earth's radiative balance by scattering and absorbing sunlight (direct effect) and acting as cloud condensation nuclei (indirect effect), thereby influencing cloud properties and lifetimes. Both moisture and aerosols influence the vertical and horizontal temperature gradients, affecting large-scale circulation patterns (Uma *et al.*, 2014). Thus, understanding the composition of the UTLS region is crucial.

Several studies in the past years have used in situ measurements, satellite observations, and modelling approaches to investigate the source regions of aerosols, water vapour, and trace gases in the UT region and the processes that control their atmospheric distribution. The Indian monsoon region is reported as one of the hotspot regions for UTLS exchange, which is often discussed as the 'Tibetan Pipe effect' (He *et al.*, 2019) or 'the monsoon air pollution pump' (Lelieveld *et al.*, 2018). Many studies have reported elevated concentrations of aerosols and other trace gases in the UTLS region during the monsoon (Alladi *et al.*, 2024; Pan *et al.*, 2016; Vernier *et al.*, 2018). Clemens *et al.* (2024) quantified that a significant fraction of continental air mass entering the UTLS region over South Asia originates from the Indo-Gangetic Plains (~30%), with additional contributions from the Himalayan foothills and the Tibetan Plateau (TP); consistent with monsoon-season studies that identified aerosol tracers originating from the Indian subcontinent in the UTLS (Clemens *et al.*, 2023).

Deep convection, such as that associated with tropical thunderstorms, mesoscale convective systems (MCS), and intense monsoon activity, is an effective

process for transporting surface water vapour, pollutants, and trace gases to the UTLS region (Singh *et al.*, 2021, 2026; Singh & Ahrens, 2025). Although there is no clear evidence that pollutants or trace gases frequently reach the UTLS directly through deep convective events, there are several studies indicating that pollutants or trace gases can reach the upper troposphere (300 hPa or above) due to deep convective events (Clemens *et al.*, 2023). Then, these pollutants or trace gases are eventually transported into the LS by large-scale phenomena such as slow transport due to slow diabatic heating, isentropic transport, and Rossby wave breaking (Clemens *et al.*, 2023, Bönisch *et al.*, 2009).

The Third Pole region has an average elevation of more than 4 km, featuring the TP and the high and complex Himalayan mountain ranges to the south (Kukulies *et al.*, 2023; Singh & Ahrens, 2025). It also plays an important role in synoptic phenomena such as monsoon circulations. Therefore, understanding the Third Pole's role in the UTLS transport is essential. Previous studies show that MCSs are frequently observed in the Third Pole region, especially during the Indian summer monsoon (Kukulies *et al.*, 2023), whereas the frequency of overshooting deep convection is low in this region compared to the tropics (Huang *et al.*, 2023). Since the Third Pole region is more elevated than the other monsoon-affected areas, we have presumed that deep convection over the Third Pole still can contribute significantly to the transport of near-surface constituents in the UT and possibly into the LS.

In recent years, km-scale (≤ 4 km) horizontal resolution simulations over the Third Pole region have proven beneficial for understanding the patterns and processes associated with extreme weather events in the region (Collier *et al.*, 2024; Karki *et al.*, 2018; Kukulies *et al.*, 2023; Prein *et al.*, 2022; Singh *et al.*, 2021; Singh & Ahrens, 2023, 2025). Km-scale simulations provide higher resolution and improved representation of convective processes, but still require careful treatment of shallow convection (Prein *et al.*, 2015). Some studies have shown that km-scale simulations with shallow convection parametrisation can capture extreme precipitation events in the complex terrain of the Third Pole region (Collier *et al.*, 2024; Prein *et al.*, 2022; Singh *et al.*, 2021; Singh & Ahrens, 2023, 2025), although these investigations were

limited to individual events. Previous studies of water vapour and pollutant transport into the UTLS over the Indian subcontinent largely relied on ERA5/-Interim datasets, where convection was parametrised (Clemens *et al.*, 2023; Hanumanthu *et al.*, 2020). Consequently, there is a strong possibility that the role of convection in determining the trajectories of pollutants from the source regions to the UT and LS may be misrepresented.

Therefore, this study aims to understand the pathways and processes of water vapour transport to the UT in the Third Pole region during extreme events. With the help of various simulations, we tried to quantify the impact of shallow and deep convection parametrisation in the vertical transport of water vapour to the UT region during extreme events. This manuscript consists of four sections. The first section serves as an introduction. The second section provides details of the various data sets used in this study, event selection, and different sets of model simulations. The third section explains the results and discusses the findings in detail. The fourth section provides a summary and conclusions of our study.

2 Data and methodology

Due to the scarcity of in situ observations over the Third Pole region, our study used various reanalyses, satellite observations, and model simulations at different spatial and temporal scales to evaluate simulations of extreme events.

2.1. Reanalysis data

In this study, we used two reanalysis datasets, ERA5 and IMDAA, for the Third Pole region. Both reanalyses datasets were produced using an atmospheric model with parameterised shallow and deep convection.

2.1.1. ERA5

The fifth generation of ECMWF atmospheric reanalysis (ERA5) hourly data at the surface and various pressure levels were used for comparison purposes (last accessed on 20/02/2026; <https://cds.climate.copernicus.eu/>; DOI: 10.24381/CDs.adbb2d4). ERA5 provides hourly data for various atmospheric, oceanic, and land surface quantities from 1940 to the present. ERA5 atmospheric reanalysis data is available at ~ 30 km horizontal grid resolution and 137 vertical pressure levels of up to 0.01 hPa (Hersbach *et al.*, 2020).

2.1.2. IMDAA

The Indian Monsoon Data Assimilation and Analysis (IMDAA) hourly data at the surface level and various

pressure levels were used for all events (last accessed on 20/02/2026; <https://rds.ncmrwf.gov.in/>). The IMDAA atmospheric reanalysis is available for the Asian domain (15° S– 45° N, 35° E– 115° E) from 1979 onward. IMDAA is available with 12 km horizontal resolution and 63 vertical level from the surface to 40 km altitude. IMDAA-derived Indian monsoon features closely correlate with the Indian Meteorology Department's observations and ERA5 (Hunt *et al.*, 2022; Indirarani *et al.*, 2021).

2.2. Satellite data

This study leverages a range of satellite and remote sensing datasets for validation and analysis.

2.2.1. AIRS

The Atmosphere Infrared Radio Sounder (AIRS) on NASA's Aqua satellite is a multi-infrared and near-infrared channel advanced sounder used to obtain significantly accurate values of temperature and water vapour profile throughout the atmosphere. We used Level 3 AIRS data for the evaluation of water vapour and temperature (last accessed on 20/02/2026; <https://giovanni.gsfc.nasa.gov/>).

2.2.2. IMERG

Integrated Multi-satellitE Retrievals for Global Precipitation Measurement (GPM) (IMERG) used combined information from the GPM satellite constellation to estimate global precipitation. In this study, we used half-hourly at 0.1-degree spatial resolution, version 06 (GPM_3IMERGHH) Level 3 (last accessed on: 20/02/2026; https://disc.gsfc.nasa.gov/datasets/GPM_3IMERGHH_06).

2.2.3. CMORPH

Climate Prediction Center (CPC) MORPHing technique (CMORPH) derives precipitation at high resolution using observations obtained from low orbiter satellites. This study used 0.25-degree spatial resolution hourly data version V1.0 (last accessed on: 20/02/2026; <https://www.ncei.noaa.gov/products/climate-data-records/precipitation-cmorph>).

2.2.4. CALIPSO

Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) on NASA's Aura satellite measure various atmospheric parameters, including atmospheric composition, temperature, humidity, and cloud ice. In this study, we used information combined with CALIPSO lidar, the Infrared

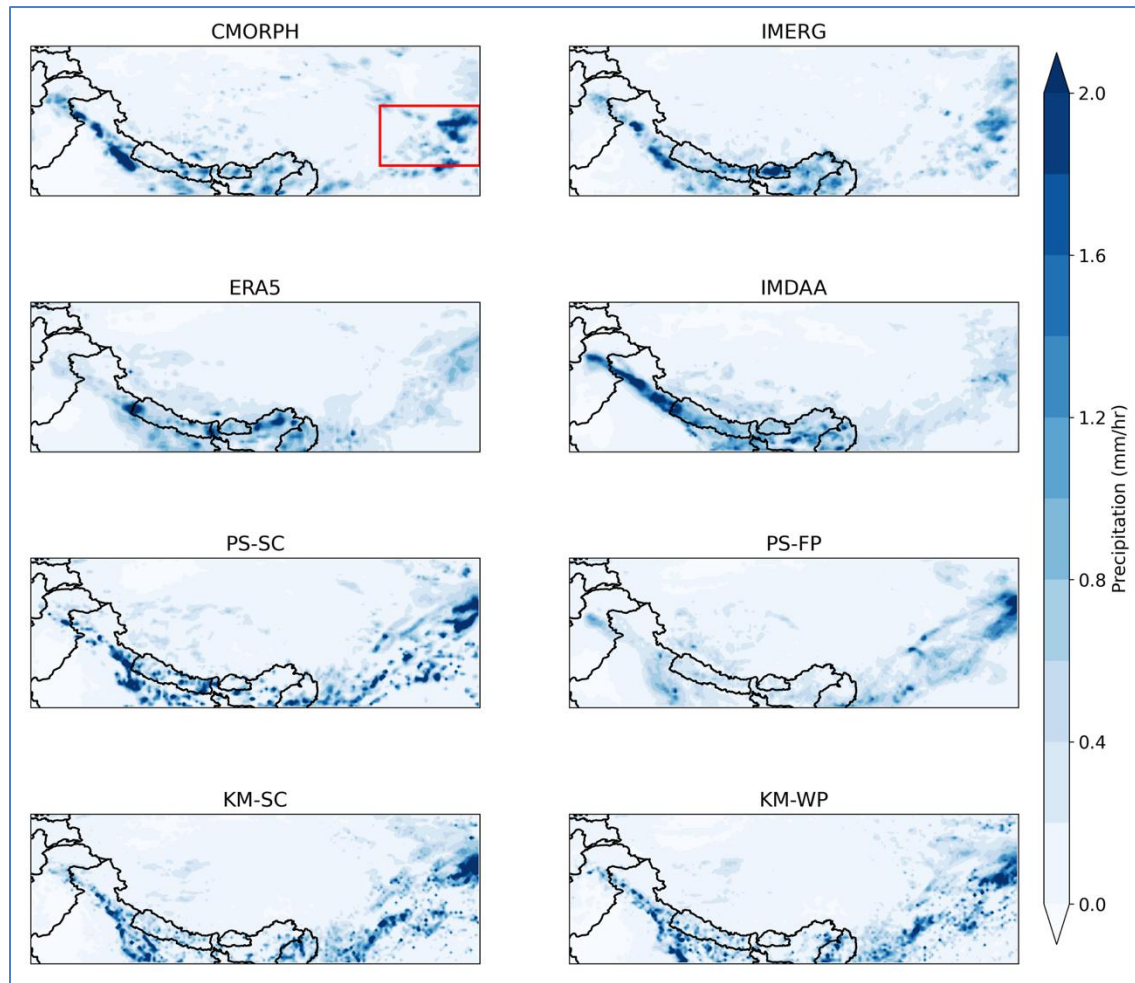


Fig. 1. Two days' average precipitation (in mm/hr) for the first event (E1) from satellite-based observations (CMORPH, IMERG), reanalysis (ERA5, IMDAA), and ICON-CLM simulations at 13 km (PS-SC, PS-FP) and 3.3 km resolution (KM-SC, KM-WP). The red box shows E1's heavy precipitation region

Imaging Radiometer (IIR), and Wide Field Camera (WFC) from CALIOP Level 1B and Level 2 data product, version 4 (4.51) (last accessed on: 20/02/2026; <https://www-calipso.larc.nasa.gov/products/lidar/browse-images/production/>). This data set contains images from a 532 nm total attenuated back-scatter, which provides a tropopause layer over the pass area. In addition, the ice/water phase, aerosol subtype, and cloud subtype are also used from CALIPSO to understand the convection activity in the study region.

2.3. Extreme events

In this study, we focus on three extreme precipitation and snow events driven by different atmospheric processes, seasons, and impacts in the Third Pole region (Prein *et al.*, 2022), over the area illustrated by the red boxes in supporting Fig. 1. The first event (referred to as E1) involved the formation of a long-lasting mesoscale TP

vortex over the east of the Third Pole in mid-July 2008 (Feng *et al.*, 2014). The second event (E2) occurred in the western part of Nepal and the eastern part of Uttarakhand state, India during monsoon, on August 14-15, 2014 (Karki *et al.*, 2018). August 2014 was noted as one of the wettest Augusts in recent decades. The third event (E3) corresponds to a heavy snow event in October 2018 at the Nam Co station (30.66° N, 90.64° E) in the south-central part of the TP driven by lake-effect (Dai *et al.*, 2020). The TP region generally experiences a low frequency of deep convection activities, with an increasing frequency from the west to the east (Huang *et al.*, 2023). In contrast, the southern side of the TP, particularly the Indo-Gangetic Plains (IGP), experiences a relatively higher frequency of deep convection activity within the study area. Consequently, the three cases represent varying strengths of deep convection activity in the domain, E1 representing the strongest convection activity region near the Third Pole, E2 in the IGP, and E3 in the south-central part of the

Third Pole, where deep convection is comparatively infrequent (Devasthale & Fueglistaler, 2010). These events are extensively discussed within the convection-permitting Third Pole (CPTP) - COordinated Regional Downscaling EXperiment (CORDEX) flagship project (Prein *et al.*, 2022).

2.4. Model setup

In this study, we have used the Icosahedral Nonhydrostatic (ICON) weather and climate model in Climate Limited-area Mode (CLM; ICON-CLM), version 2.6.5 (Van Pham *et al.*, 2021). One week spin-up was considered for every event we simulated over the Third Pole region (TPR). The E1 case study period was 14th – 24th July 2008, the E2 case study period was 27th July to 1st September 2014, and the E3 case study period was 4th – 9th October 2018. The ICON-CLM simulation was performed at two resolutions for each event, Convection Permitting or km-scale (3.3 km) and Convection Parametrized scale (13 km) with 50 vertical levels up to 10 hPa covering the domain 22.5° N – 42.5° N and 67.5° E – 117.5° E. We have used a smaller subdomain (25° N – 40° N and 70° E – 115° E; Supporting Fig. 1) for our analysis. As well as only 21 pressure levels from 1000 hPa to 10 hPa have been used for output storage. Initial and lateral boundary conditions were provided through the ERA5 data set. The lateral boundaries were updated every hour during the simulations.

For all the simulations, we have used the ecRad radiation scheme (Hogan & Bozzo, 2016), the same setup as used in the previous ICON-CLM simulation for the CPTP CORDEX (Collier *et al.*, 2024; Singh *et al.*, 2026; Singh & Ahrens, 2023, 2025). The Tiedtke-based shallow or deep convection parametrization was enabled whenever it was needed in the simulations (Tiedtke, 1989). The land surface process was considered through the sub-model TERRA (Schulz *et al.*, 2016) and single-moment cloud microphysics (Doms *et al.*, 2021), other settings were the same as the previous study over the TPR using ICON-CLM (Collier *et al.*, 2024; Singh & Ahrens, 2023, 2025).

2.4.1. Convection permitting scale (km-scale)

At km-scale (1–5 km of grid spacings) simulations explicitly treat moist convection. Some experts treat the 1 – 5 km scale as a grey zone: completely turning off convection parametrization can be risky on this scale (Ahrens *et al.*, 2003; McTaggart-Cowan *et al.*, 2019; Prein *et al.*, 2022; Tomassini *et al.*, 2023). Some studies on the TP and the Himalayan Mountains suggest using only shallow convection parametrization without the deep convection parameterisation at km-scale (Singh *et al.*, 2021; Singh & Ahrens, 2023, 2025). Therefore, with the

TABLE 1

Summary of the model set-up for ICON-CLM simulations

Simulations name and type	Grid Space	Convection Parametrization		Simulation time step
		Deep	Shallow	
KM-SC	3.3 km	OFF	ON	25 Sec
KM-WP	3.3 km	OFF	OFF	25 Sec
PS-SC	13 km	OFF	ON	60 Sec
PS-FP	13 km	ON	OFF	60 Sec

above-specified ICON-CLM setup, we did two sets of simulations for each event at km-scale: one km-scale with only shallow convection parametrization (KM-SC) and one without any parametrization of convection (KM-WP).

2.4.2. Convection Parametrized scale

Horizontal resolution below 5 km is a scale where we must parametrise the moist convection (McTaggart-Cowan *et al.*, 2019). Therefore, we also did two simulations for each case, one with only shallow convection parametrization (PS-SC) and one with deep convection parametrization (PS-FP) at 13 km grid spacing. Switching of the deep convection parameterisation is motivated by Bougeault and Geleyn, (1989) who showed that grid-scale convection is partly able to compensate for lack of deep convection parameterisation even at this coarse grid scale.

We now have four sets of simulations for each extreme event case (Table 1), which we compared with the ERA5 and IMDAA reanalysis data, both of reanalysis also use convection parametrization.

2.4.3. Off-line Lagrangian Trajectory Model

In this study, we used a simplified Lagrangian trajectory model to calculate parcel trajectories based on wind fields from all simulations and ERA5 for all extreme events. This offline model considers the initial position of the parcel and integrates parcel movement over time (Hernández Pardo, 2023, 2024; Miltenberger *et al.*, 2013). It does not parametrise vertical winds to compensate for deep convection; therefore, it represents deep convective transport directly from its driving dataset without any adjustment. This code can be easily accessible through its git repository (last accessed on: 20/02/2026; https://github.com/lhpardo90/trajectory_calculator.git).

We performed both forward and backward trajectory calculations with the Lagrangian model to track air parcels during all extreme events. For each event, the trajectories of 50 air parcels were calculated every hour, both forward and backward, throughout the event period. These parcels

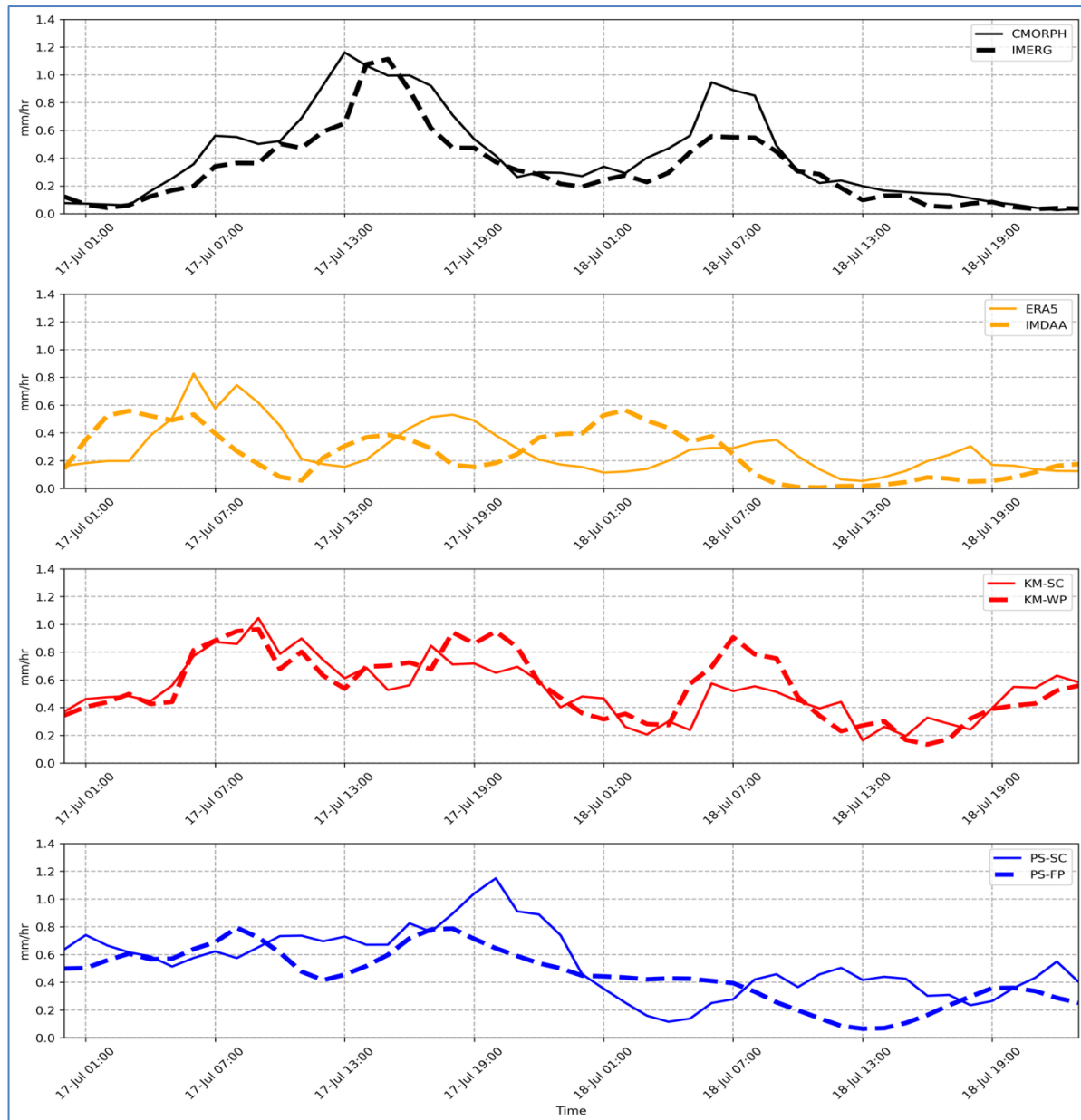


Fig. 2. Two-day hourly precipitation average over the first event region (E1) from observations (black), reanalysis (orange), and ICON-CLM simulations at km-scale resolutions (red) and 13 km resolutions (blue)

were placed at different points and pressure levels (near the surface to 700 hPa) within the domain of each extreme event case.

3. Result and discussion

As discussed earlier, in this study we used four model configurations at different resolutions along with ERA5 and IMDAA reanalysis. Therefore, examining the representation of orography in these datasets provides a basis for understanding subsequent model evaluation and performance.

3.1. Resolution Dependence of Orography and Process Parameterisation

High-resolution model simulations offer several key advantages, including better resolution of topographical features compared to low-resolution model simulations (Brisson *et al.*, 2017). The features of mountains and valleys are better captured at km-scale resolution compared with low-resolution models like ICON-CLM at 13 km resolution and ERA5. For example, the discrepancy is evident when examining the elevation of Mount Everest (27.98 °N, 86.92 °E), which is ~ 8849 m; in the km-scale

simulation using ICON-CLM, the elevation is 7207 m, while the 13 km simulation yields 6260 m, and ERA5 shows 5491 m at their respective grid resolutions. The valleys are also well represented in high-resolution models; previous studies using high-resolution (1 km) WRF simulation in the central Himalayan region showed quite good agreement with the elevation of the observation stations, while 5 km and 25 km showed an overestimation of 1-1.5 km (Collier & Immerzeel, 2015; Karki *et al.*, 2017). This disparity highlights the capacity of high-resolution models to better represent topography, with significant implications for various localised atmospheric processes such as katabatic winds, seasonal changes, topographical interaction of winds, *etc.* In addition, high resolution reduces the need to parametrise certain small-scale processes such as convection or gravity waves (Prein *et al.*, 2015, 2022; Singh & Ahrens, 2023; Zhou *et al.*, 2021).

3.2. Evaluation of Simulated Extreme Events

Precipitation from different simulations shows vast spatial and temporal variability and strong sensitivity to the choice of convection parametrisation in simulations. On the other hand, 2m temperature does not show strong spatial-temporal differences across the datasets in the TP region.

Fig. 1 shows mean precipitation in the domain for event E1. CMORPH shows relatively high precipitation (> 2 mm/hr) over the event region. IMERG also captures the high precipitation (> 1.2 mm/hr) over the same region but not as high as CMORPH. A comparative study between gauge-based observed precipitation and satellite-based observations has reported that the IMERG records are lower than gauge observations and CMORPH over the TP (Wei *et al.*, 2018). The ERA5 and IMDAA reanalysis captures high precipitation in the region, but less than observed. In contrast, in the IGP and foothills of the Himalayas, ERA5 and IMDAA are comparable to satellite observations. The coarse ICON-CLM simulation with deep convection parametrisation (PS-FP) captures features similar to those of reanalyses and satellite. The coarse PS-SC represents slightly higher precipitation and closer to satellite observations than PS-FP. The km-scale ICON-CLM simulations KM-SC and KM-WP compares best to satellite observations. The flagship study CPTP-CORDEX has reported that ICON-CLM simulation at km-scale simulation with shallow convection parametrisation (KM-SC) captures the spatial features of event E1 better than many of the participating models (Prein *et al.*, 2022).

The temporal analysis of satellite observations from CMORPH and IMERG shows a high precipitation rate in the E1 region between 12 – 18 UTC on 17th July and 6 –

12 UTC on 18th July (Fig. 2). Both observations capture the first precipitation peak with similar strength, whereas IMERG shows lower precipitation during the second peak. ERA5 and IMDAA show lower precipitation than satellite observations. Both reanalyses could not reproduce any precipitation peak. IMDAA could not produce strong precipitation in the E1 region. Both reanalyses seem to follow a precipitation peak similar to the diurnal peak of convective available potential energy (CAPE) in the E1 region. Both ICON-CLM simulations at 13 km resolution (PS-SC and PS-FP) show higher precipitation than the reanalyses data in the E1 region. PS-FP captures the first high precipitation peak, with some time lag, and completely misses the second peak. The PS-SC simulation captures both peaks, but with some time lag. The km-scale simulations KM-SC and KM-WP simulated higher precipitation values in the E1 region than the observations. They captured both observed peaks and outperformed ERA5 and low-resolution ICON-CLM simulations in terms of spatial and temporal features and precipitation intensity. The KM-WP simulation, *i.e.* without shallow convection parametrisation, shows slightly higher precipitation values than the KM-SC.

All ICON-CLM simulations show a similar trend in CAPE within the E1 region. The CAPE peaks around 6 UTC with comparable intensity across simulations, while PS-FP exhibits the lowest CAPE values on both days (Supporting Fig. 4). The precipitation peak of ERA5 and PS-FP corresponds to CAPE peak. Meanwhile, PS-SC shows a delayed precipitation peak. The precipitation peaks from the km-scale simulations were near the CAPE peaks, but with slower decay than in ERA5. Km-scale simulations had CAPE magnitudes similar to those of ERA5, but the magnitude of the precipitation peaks were closer to the satellite's observed values than ERA5's. A previous study by Hohenegger *et al.*, (2008), conducted using COSMO-CLM with similar physics parameterisations as in ICON-CLM, suggests that km-scale simulations can better reproduce precipitation at both spatial and temporal scales. Additionally, the study highlighted that km-scale simulations exhibit a more realistic delay in the convective precipitation peak and a longer decay time compared to coarser resolution simulations. ERA5 and PS-FP show a larger wet area than other simulations and satellite observations. This indicates that despite having similar atmospheric instability, the models at different resolutions and convection parametrisation choices have different responses in representing the convective precipitation.

For E2, CMORPH and IMERG both show higher precipitation starting at noon on 14th August and lasting until the early morning of 15th August (Supporting Fig. 2). IMERG shows a peak at ~ 22 UTC which matches the

precipitation peak observed from the automated weather station (Karki *et al.*, 2018). The IMDAA reanalysis captured this event and its duration, with some differences in the precipitation peak compared to IMERG. The ERA5 reanalysis underperformed for E2: spatial and temporal features were completely missing. Whereas ICON-CLM's low-resolution simulation PS-SC captured this event quite well compared with CMORPH but underestimated the peak magnitude, PS-FP with deep convection parametrisation missed this event completely, same as ERA5. The km-scale simulations captured the spatial features of precipitation with some difference on the temporal scale compared to the satellite observations. The intensity of all ICON-CLM simulations (except PS-FP) was comparable to the observations but higher than that of ERA5. All models have CAPE peaks with similar timing on both days of E2. ERA5 and PS-FP had high CAPE values (800 J/kg); both showed larger wet areas during E2 compared to other simulations and satellite observations. Synoptic features indicate that the northwestern monsoon flow converges with upper-level westerlies over the E2 region, a pattern consistently observed across all simulations and previously reported by (Karki *et al.*, 2018). Studies have highlighted that the interaction between monsoon flow and mid-latitude westerlies in the foothills of the Himalayas is an important factor that contributes to flooding in the region during the monsoon season (Singh *et al.*, 2021, Priya *et al.*, 2016).

For the snow event E3, the ERA5 reanalysis shows more precipitation than the satellite observations for all five days. Satellite observations show a diurnal pattern, leading to the precipitation peak in the afternoons on all days (Supporting Fig. 3). The low-resolution ICON-CLM simulations match the satellite observation, but km-scale simulations underperformed for E3. A previous study by Dai *et al.* (2020) reported that the 4th–8th October extreme snow case was one of the strongest observed extreme snow events in the last decade. They reported that the east and south of the Nam Co station faced heavy snow on 8th October, which is closely simulated in all ICON-CLM simulations with respect to satellite observations with some spatio-temporal differences. ERA5 showed higher CAPE than ICON-CLM simulations; diurnal peaks of precipitation were quite close to the diurnal peaks of CAPE. ERA5 shows a larger wet area than the ICON-CLM simulations and satellite observations for all days of E3. Compared with E1 and E2, E3 exhibits the lowest precipitation amounts and CAPE values (Supporting Fig. 4). This is likely related to the dominant influence of the lake-induced mesocirculation during this event. Previous studies have shown that, in the absence of sufficient moisture and hydrometeor sources, lake-induced circulation over the Tibetan Plateau primarily generates shallow ascent and weak convection, limiting the

development of deep convective clouds (Yang *et al.*, 2026; Zhang *et al.*, 2024). Consequently, the lake-induced circulation alone appears insufficient to produce strong deep convection over the plateau, which is consistent with the relatively low precipitation and CAPE observed during E3.

For the three extreme events, the km-scale ICON-CLM precipitation simulations exhibit an overall higher correlation with observations than both the reanalyses and the low-resolution simulations (Fig. 3), with KM-SC on average better than KM-WP. The correlation values were smaller in the case of the snow event E3 than for the other events, with the correlation of the satellite products about zero. It is interesting to note that the coarse-grid ICON-CLM simulation with parametrised deep convection performs worse than without. Overall, the ICON-CLM simulations correlate stronger with ERA5 (the forcing data of the simulations) than with IMDAA. The 2-meter temperature cross-correlations were high (> 0.9) except for the km-scale simulation with parametrised shallow convection in the snow event E3 (between 0.84 and 0.88).

3.3. Water vapour in the upper troposphere

Given the results discussed in the former sub-section, we expect the km-scale simulations to represent a better water vapour profile during extreme events than low-resolution simulations over the Third pole region.

The AIRS observation shows high values of specific humidity at 100 hPa for E1 (Fig. 4), which are overestimated by all ICON-CLM simulations and less so in ERA5. But ERA5 has the highest values in other parts of the analysis domain. Similarly, for E2, AIRS shows slightly high specific humidity near the region of interest, which is simulated quite well in all simulations, and ERA5. Unlike E1, for E2, km-scale and low-resolution ICON-CLM simulations show less specific humidity than ERA5 at 100 hPa. In E3, the AIRS shows less specific humidity over the area of interest. In contrast, we can see a higher specific humidity north of the TP region than south of the TP region. This feature is quite evident in all the simulations and in ERA5. Previous studies on TP using AIRS Level-3 monthly data and long-term radiosonde observations show significant correlation coefficients for water vapour observations in the region of interest (~ 0.8). That study also highlights the main differences between both observations above 250 hPa (Zhang *et al.*, 2012). Considering these uncertainties and limitations further, we have compared simulations and ERA5.

Fig. 5 shows the time series of the area average specific humidity at 100 hPa over the each event regions

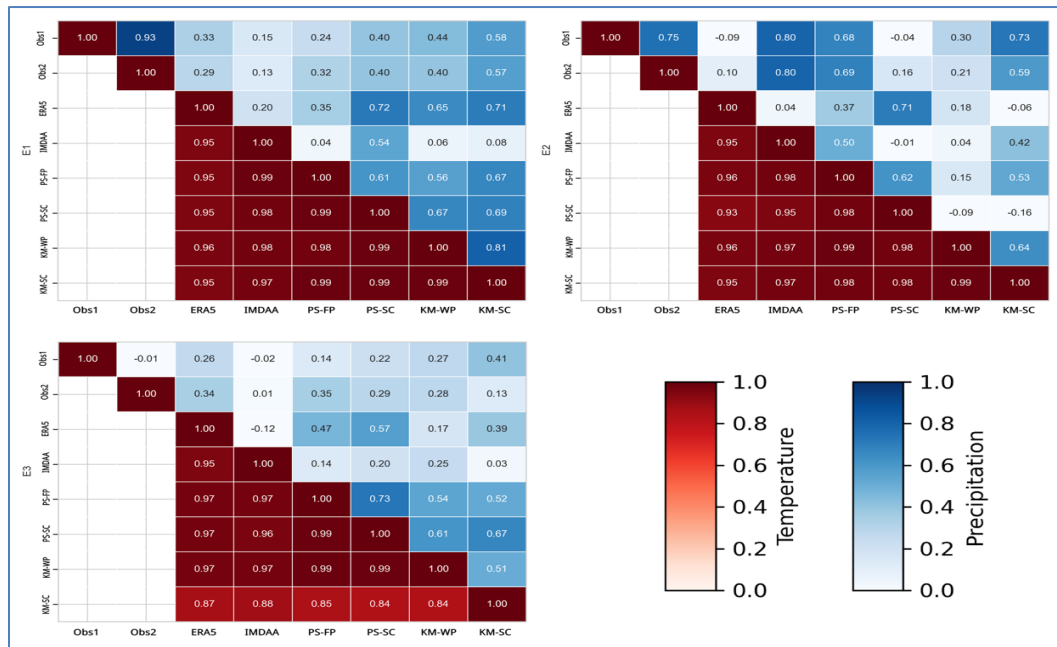


Fig. 3. Pearson correlation coefficient between satellite-based observations, reanalyses, and ICON-CLM simulations area averaged hourly precipitation and 2-meter temperature time series over all events E1, E2, and E3. The areas of averaging are indicated in Figure 4

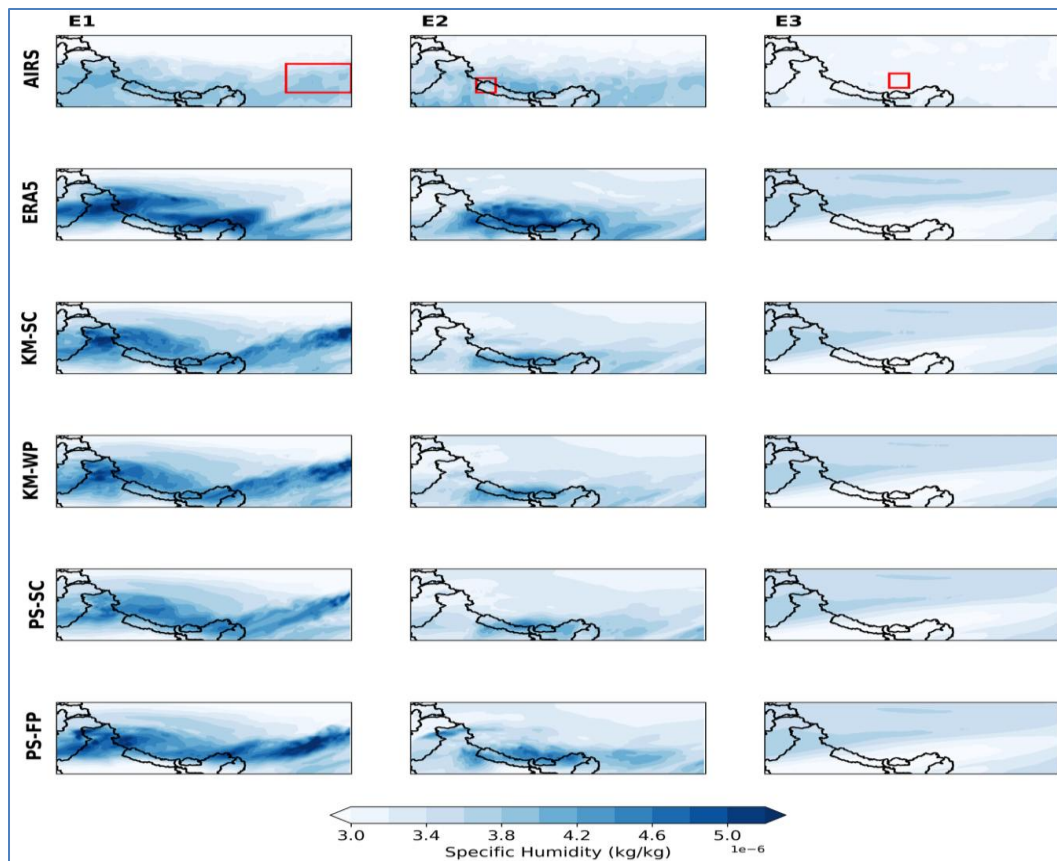


Fig. 4. Mean specific humidities at 100 hPa from AIRS observation, ERA5, ICON-CLM simulations KM-SC, KM-WP, PS-SC, and PS-FP (top to bottom) for the events E1, E2, and E3 (left to right), respectively

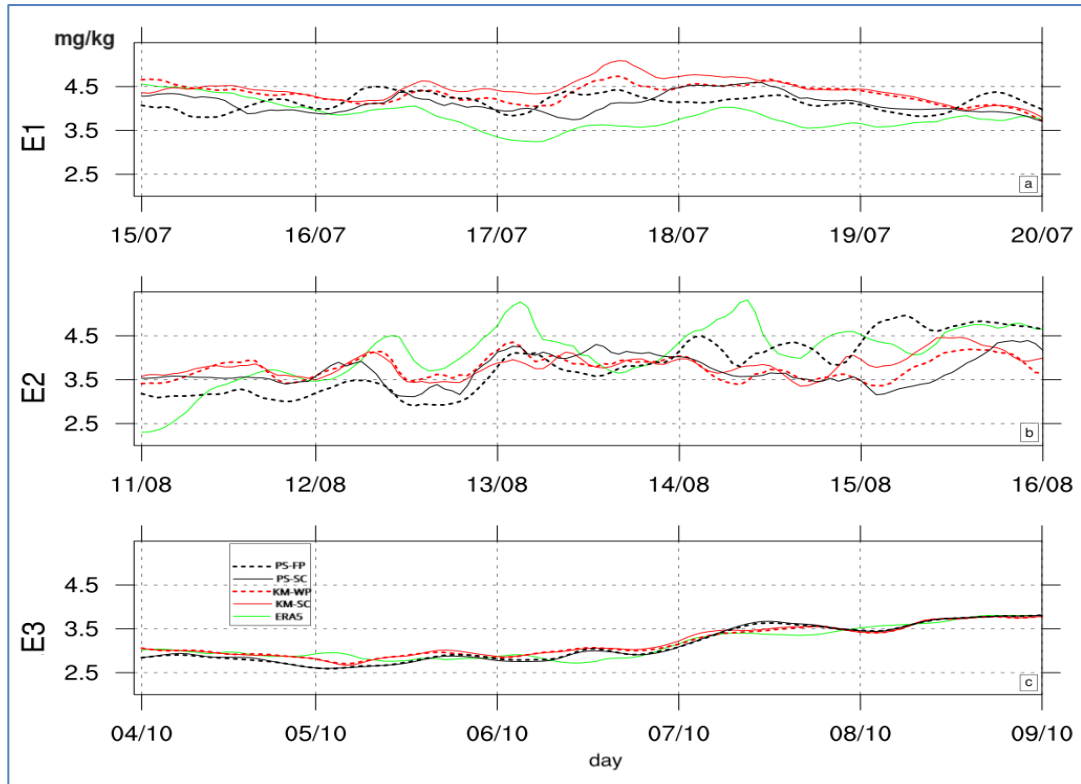


Fig. 5. Specific humidity at 100 hPa from ERA5 (green) and ICON-CLM simulations averaged over the event areas of the events E1, E2, and E3, respectively

(red boxes in Fig. 4). For E1, the km-scale simulations consistently show higher specific humidity (especially KM-SC) than the coarser-resolution simulations and ERA5 at 100 hPa. During the analysis period, KM-SC exhibits the highest specific humidity between 17–18 July compared to other days. For E2, ERA5 shows higher specific humidity than the ICON-CLM simulations, with a distinct peak present in ERA5 on the first day (14/08) of the observed event. For E3, all the ICON-CLM simulations and ERA5 show similar specific humidity at 100 hPa, with a slight increase on the fourth day of the event. This indicates that the elevated specific humidity values at 100 hPa in ERA5 are inconsistent with the deep convective events. In contrast, the ICON-CLM km-scale simulations exhibit more consistent behaviour across all events (except PS-FP for E2), with some differences.

The spatial analysis of E1 and E2 suggests an increase in specific humidity in the UTLS region due to convection (Fig. 4). E3 does not show clear evidence of increased specific humidity in any of the simulations. Meanwhile, time series analysis suggests that for E1, km-scale simulations show a higher concentration in UTLS over the extreme event region. In contrast, for E2, ERA5, and parametrised simulations show more water in the UTLS. For E3, all simulations and ERA5 show similar

concentrations in the UTLS region. If we summarise Figs. 4 and 5, the role of convection resolving simulation and deep convection in the exchange of water vapour to the UTLS region is unclear.

3.4. Convective transport of water vapour

Building on the previous subsection, we used water vapour flux over the area of observed extreme events to understand the role of the vertical updraft of water vapour due to convection. The area-averaged water vapour flux density (F_{qv}) is calculated using

$$F_{qv} = \rho \cdot qv \cdot |v| \quad (2)$$

where, ρ = air density, qv = specific humidity, $|v|$ = magnitude of the 3D wind vector

Vertical fluxes at the tropopause (~ 100 hPa) were < 0.03 mg/m²s, whereas water vapour flux values of ~ 0.5 mg/m²s were in the upper troposphere during E1 and E2 (Fig. 6). For E3, water vapour flux values were lower. In ERA5, E1 and E2 shows high total water vapour flux (~ 15 mg/m²s) up to 100 hPa, while E3 shows persistently elevated water vapour flux near the tropopause (100 hPa) throughout event period (Supporting Fig. 5). This

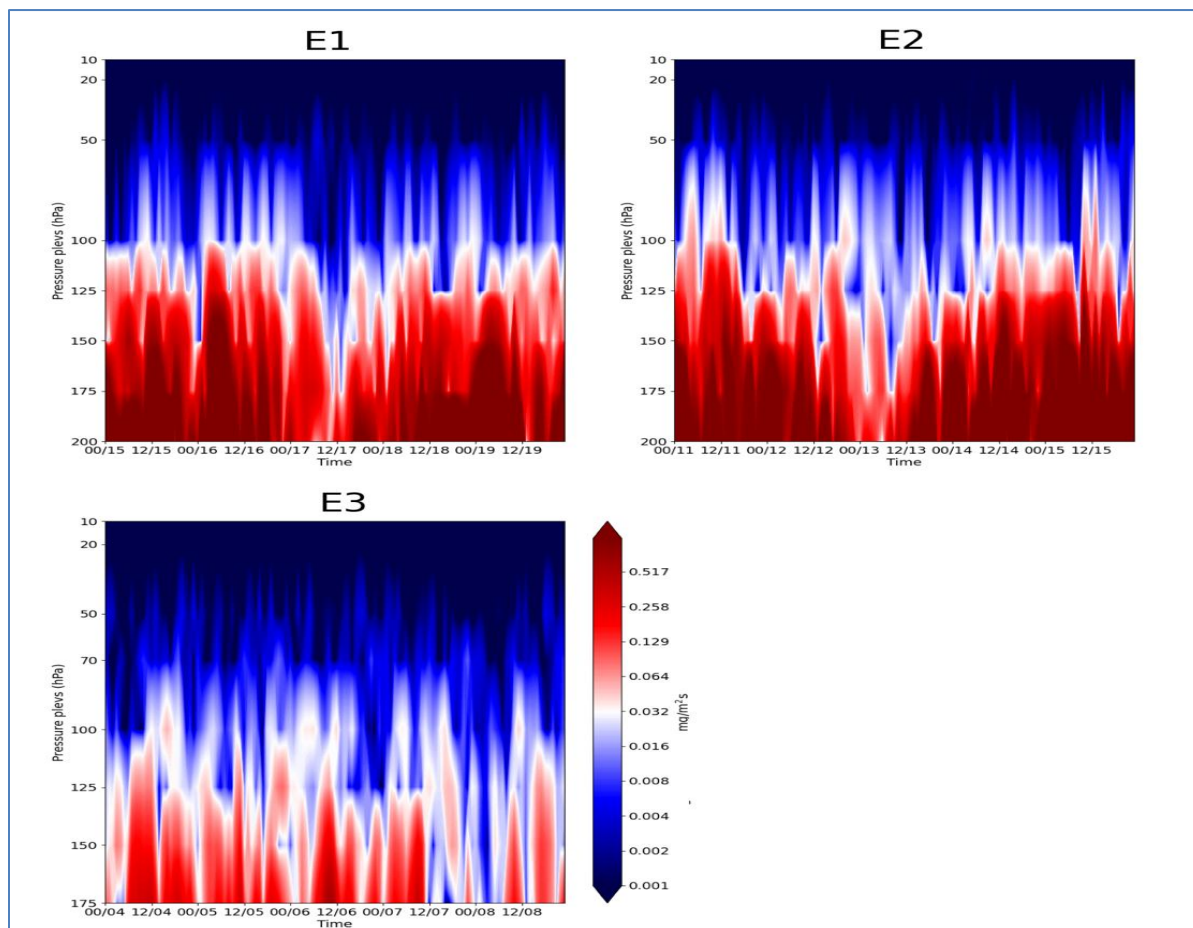


Fig. 6. Vertical water vapour flux for all three events averaged over the selected region of extreme events in ERA5

indicates that more than 95% of the water vapour flux is contributed by horizontal advection during the extreme events. Ploeger *et al.*, (2013) using the CLaMS model reported that the UTLS in the Northern Hemisphere is strongly affected by horizontal advection of moisture below 370 K originating from Asian and American monsoon systems. The TP region was mainly influenced by horizontal advection in the Asian Summer Monsoon Anticyclone (ASMA) region. Previous satellite observation and tracer-based studies over the Tibetan Plateau have further shown that air parcels reaching the upper troposphere tend to move eastward by following the subtropical jet (Honomichl & Pan, 2020; Kumar & Ratnam, 2021). ERA5 uses convection parametrisation, which may yield underestimation of vertical fluxes over the extremes' regions. Therefore, further analysis of higher-resolution simulations is presented below to clarify the role of deep convection over the Third Pole region for the UT water vapour transport.

For E1, the km-scale simulations show > 0.03 $\text{mg/m}^2\text{s}$ vertical water vapour flux above 100 hPa, slightly

higher than ERA5 (Fig. 7). The low-resolution simulation with shallow convection parametrisation (PS-SC) also shows higher flux to the UT region than ERA5 and PS-FP. The km-scale simulations show > 30 $\text{mg/m}^2\text{s}$ total water vapour flux near 100 hPa during the event period (Supporting Fig. 6). At the same time, lower resolution ICON-CLM simulations show ~ 10 $\text{mg/m}^2\text{s}$ near 100 hPa, same as ERA5.

For E2, which occurred in the central Himalayan mountains, the km-scale simulations also show higher vertical flux into the UT compared to coarser resolution, but values are much less than of total water vapour flux. The km-scale simulations show ~ 20 $\text{mg/m}^2\text{s}$ total water vapour flux above the tropopause region during the first day of the event, same as ERA5. Whereas advective water vapour flux of more than 65 $\text{mg/m}^2\text{s}$ was simulated upto 100 hPa during the period of E2, and the parameterised-scale simulations also show similar flux above the tropopause (100 hPa). Since the time of precipitation (convection) peak for event E2 varies in the km-scale, the parameterised scale, and reanalysis, the pattern of water

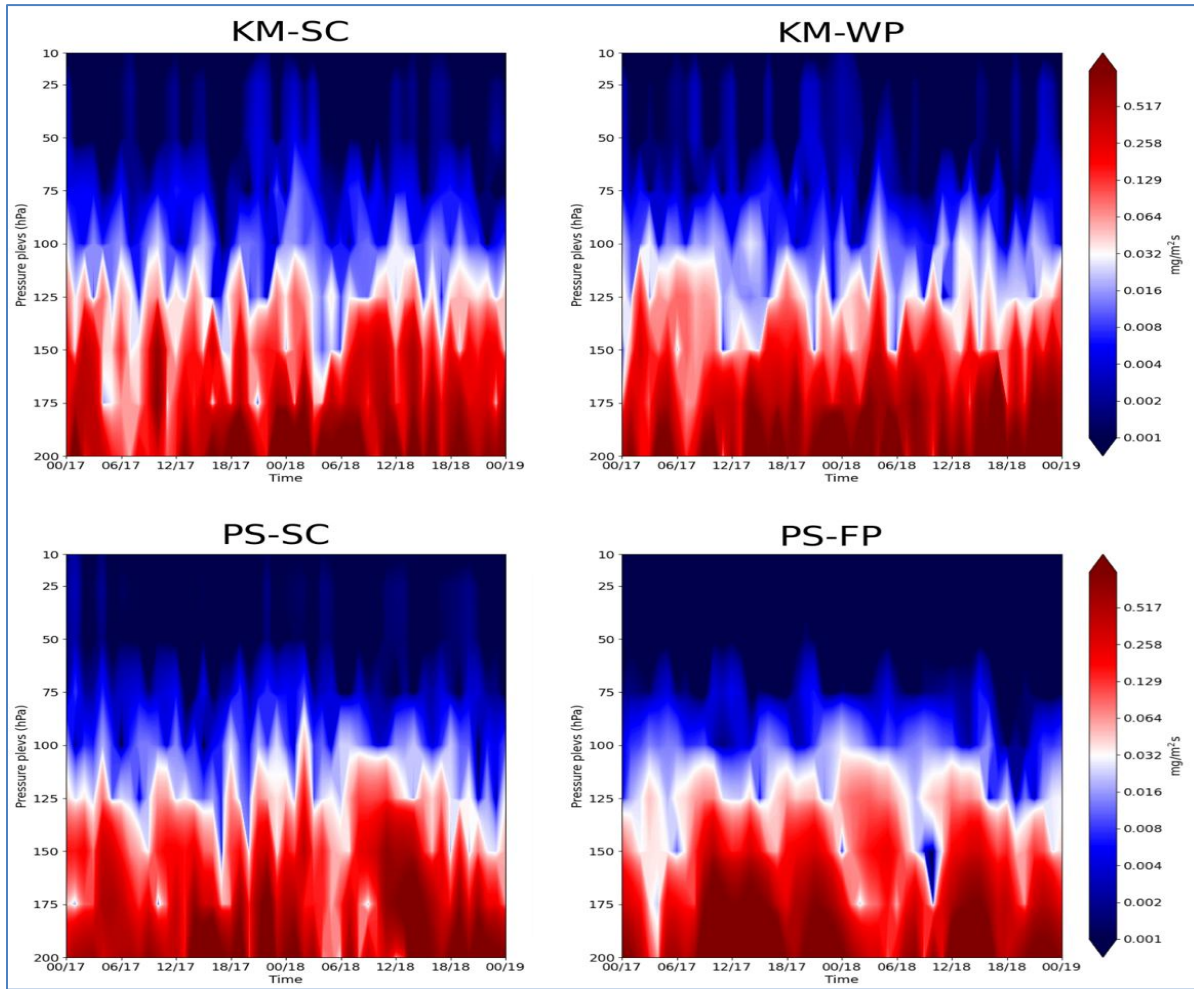


Fig. 7. Vertical water vapour flux averaged over the region of event E1 in KM-SC, KM-WP, PS-SC, and PS-FP, respectively

vapour flux also varies with respect to the time of the event. The parametrised scale simulations show more water vapour flux up to 150 hPa during the second day of the event compared to the first day. All simulations and ERA5 for E1 and E2 show very little water vapour flux in the lower stratosphere (above 100 hPa) during events, but high water vapour flux in the upper troposphere region. Horizontal advection component of the water vapour flux shows a greater influence on the UTLS water vapour than the vertical component during the monsoon season in the Third Pole region (Ploeger *et al.*, 2013).

For E3, the water vapour flux due to the vertical wind component above 100 hPa is less than 5% of the total flux above 100 hPa, similar to the other two extreme events. All ICON-CLM simulations show less total water vapour flux above the tropopause region than ERA5. However, all four simulations show water vapour flux comparable to ERA5 up to the upper troposphere. Unlike ERA5, all the simulations show increased precipitation

intensity on the fourth day, which is clearly reflected in the water vapour flux up to 100 hPa by all simulations. Like ERA5, a diurnal cycle of water vapour flux is present in all the simulations for E3, similar to the diurnal cycle of CAPE over the region, peaking in the late afternoon. During these CAPE peaks, the vertical water vapour flux is sometimes considerably high up to the tropopause.

We compared specific humidity from all simulations and ERA5 with CALIPSO observed tropopause height, ice/water phase, and cloud types, near the event regions. Fig. 8 shows one of the CALIPSO passes near E1 at 19:15 UTC on July 17, 2008. CALIPSO shows tropopause height at ~ 17 km (100 hPa) near the E1 region. In the same area, the observed 350 k potential temperature height was about ~ 9 km (300 hPa). Some deep convective clouds and water content reaching up to 12 km (300 hPa) are also observed with CALIPSO. The specific humidity from the simulations along the CALIPSO path from point 27.83° N, 99.62° E to 39.67° N, 103.04° E is shown in

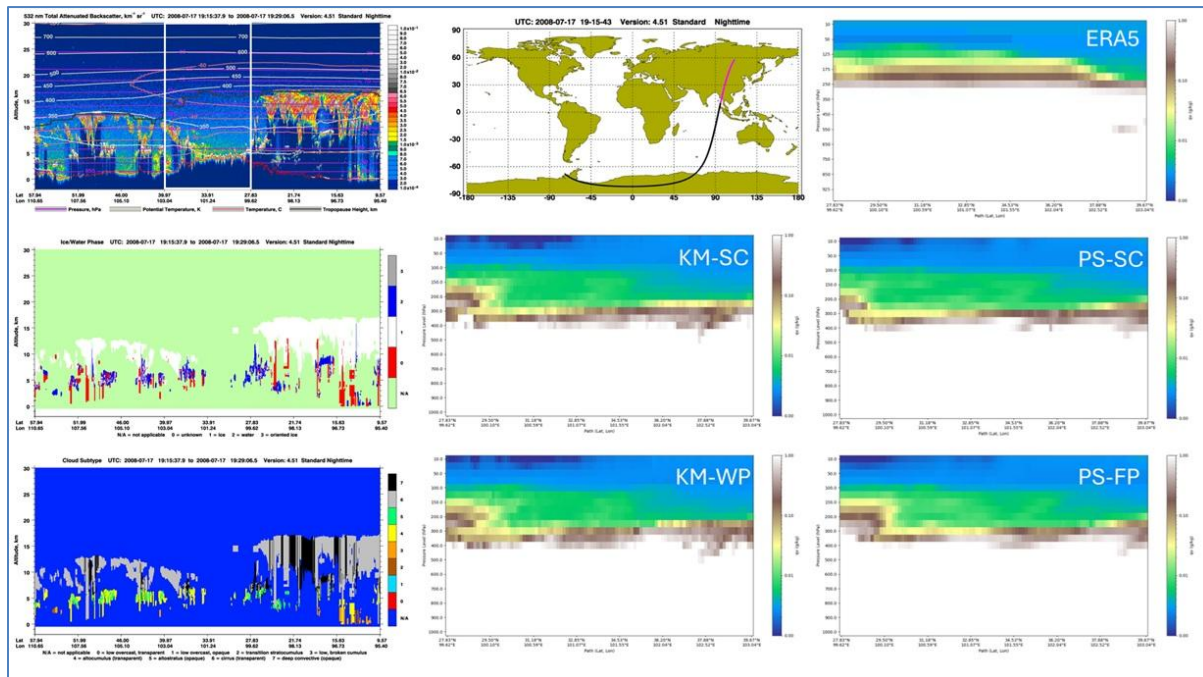


Fig. 8. The CALIPSO pass near E1 shows 532 m total attenuated backscatter-derived tropopause height, ice/water phase, and cloud subtype, along with specific humidity over E1 for the closest time stamp from KM-SC, KM-WP, PS-SC, PS-FP, and ERA5

Fig. 8. The km-scale simulations show elevated specific humidity up to 150 hPa, similar to CALIPSO water/ice phase observation. The specific humidity has increased near 100 hPa in the observed deep convective cloud region from CALIPSO (27.83° N, 99.62° E) than stable regions. The parametrised scale simulations also closely follow the features of the km-scale simulations. ERA5 shows a higher specific humidity at the northern edge of the CALIPSO pass and lower at the observed convective region (27.83° N, 99.62° E), different from the CALIPSO observation. Overall, all ICON-CLM simulations agree with CALIPSO observations and show a moist upper troposphere (300–200 hPa) above the 350 K potential temperature line for E1.

For E2, three CALIPSO passes were available, and for E3, six CALIPSO passes were available in the region and duration of extreme events. All CALIPSO passes for E2 show tropopause height ~ 100 hPa (~ 17 km), and the 350 K potential temperature line was ~ 300 hPa (~ 10 km). The convective clouds and the water/ice phase derived from CALIPSO reach up to 150 hPa (15 km), very close to the tropopause (100 hPa), and well above 350 K potential temperature line (300 hPa). Similarly, simulated specific humidity along all CALIPSO passes show a higher values, reaching up to 150 hPa, over the deep convective regions (observed by CALIPSO) compared to stable regions. ERA5 shows slightly higher values near 100 hPa compared to the simulations. Similarly, for E3,

all the CALIPSO passes show tropopause height ~ 100 hPa (16–17 km) and 350 K potential temperature line ~ 200 hPa (12–13 km). In most cases, CALIPSO indicates deep convective clouds and water/ice reached above 200 hPa (~ 12 km). All simulations show moist upper troposphere (up to 150 hPa), and relatively more specific humidity at 100 hPa, which is not that evident in CALIPSO observations (Fig. 9). All the simulations agree with CALIPSO's deep convective signature in space and time, whereas ERA5 slightly overestimated specific humidity in spatial coverage. It's possibly because high-resolution ICON-CLM simulations better capture local processes than ERA5.

3.5. Lagrangian Tracking of Air Parcels

In addition, we used Lagrangian tracking of air parcels to assess the potential of extreme events to rapidly uplift air parcels into the UTLS. Fig. 9 shows the Lagrangian paths of air parcels for all three events, initiated during the active phases (precipitation peak) and calculated forward in time for 12 hours to capture the immediate effects of deep convection. For E1, both ERA5 and KM-SC show that many parcels reached above 300 hPa and moved eastward, but none directly reached 100 hPa. It is clear that for E1 during the event, air parcels rapidly reach a potential temperature height of up to 350 K and then move with the ASMA winds. In addition, to confirm this pattern, we initiated Lagrangian simulation

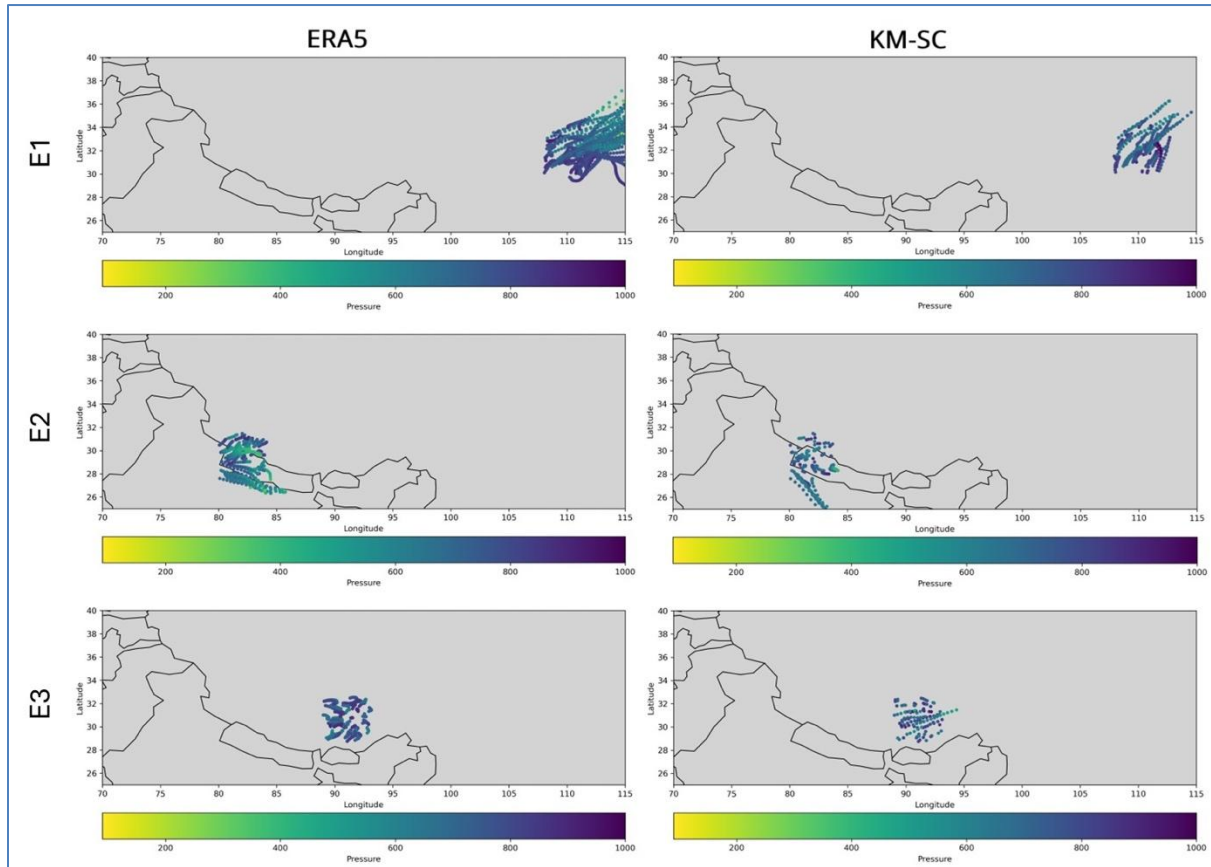


Fig. 9. Lagrangian air parcel transport path for E1, E2, and E3 initialised near the surface of respective events and traced for 12 hours in forward mode

using ERA5 and KM-SC at different initial time steps; for most of the simulations in the first few hours, only a few air parcels ascend to ~ 300 hPa and then move eastward. In E2, relatively more air parcels reached the upper troposphere. The km-scale simulations show more localised vertical transport of air parcels during events E1 and E2 compared to ERA5. E3 shows the weakest transport of the extreme events; some air parcels reached 300 hPa with ERA5, whereas KM-SC shows that very few can reach above 300 hPa.

We performed several back-trajectory analyses using ERA5 and KM-SC for all the events. The back trajectory initiated above 100 hPa shows air parcels coming horizontally from the west, the usual ASMA flow. Previous studies reported that in the active convection phase during the Asian monsoon season, air parcels were observed to reach 350 K near the subtropics, confined to the same layer and tend to move horizontally into the mid-latitude (Ploeger *et al.*, 2013; Yan *et al.*, 2021). This advection is indicated in Fig. 4 with an extremely high specific humidity observation above the western Himalayan region. Since these regions are dominated by

deep convective events during pre- and early monsoon (Singh & Ahrens, 2023, 2025), we performed several Lagrangian tracking of air parcels from ERA5 and KM-SC simulations in the western Himalayan region. The KM-SC simulation shows more air parcel transport to the upper troposphere region than E1, E2 (central Himalayan event). In contrast, ERA5 indicates a similar strength of transport over the western Himalayas as in the central Himalayan event (E2). This indicates that deep convective events over the Himalayan Mountains can lead to stronger transport to the upper troposphere, but not strong enough to cross the tropopause region.

In addition to Lagrangian tracking of air parcels, we quantified the water vapour flux along the Lagrangian path for every air parcel which reached at least 300 hPa for the three extreme events. For E1 (Fig. 10), ERA5 and the coarse ICON-CLM simulations show similar mean water vapour flux along air parcel trajectories in the lower- and mid-troposphere, but the km-scale simulation show larger fluxes. In the upper troposphere the ICON-CLM fluxes are comparable, but ERA5 shows larger fluxes.

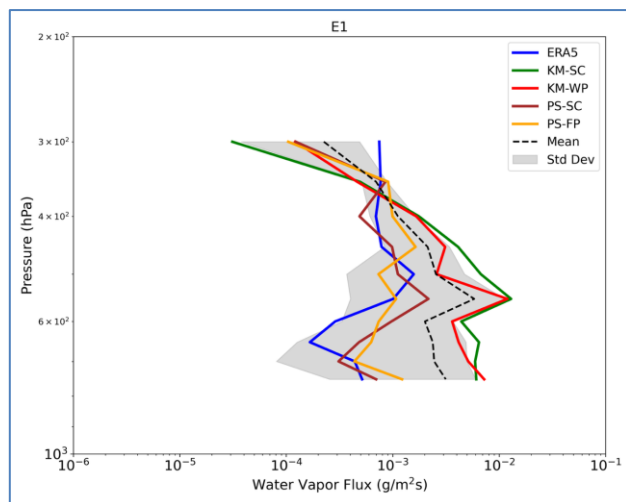


Fig. 10. Average water vapour flux along Lagrangian air parcels in forward transport path for all the E1 air parcels reaching at least 300 hPa

For E2, ERA5 and PS-FP show higher transport along trajectories, followed by KM-WP, KM-SC, and PS-SC. For E3, KM-WP shows the largest flux values, followed by PS-FP, KM-SC, PS-SC, and then ERA5, but with consistent vertical flux gradients. Overall, ERA5 appears inconsistent with respect to water flux along the Lagrangian path in the upper troposphere during ascent. A previous study using various reanalyses' data sets suggests that reanalyses overestimate the water vapour flux contributed due to the synoptic process, especially ERA5 (Deng *et al.*, 2022). The ICON-CLM simulations KM-WP, KM-SC, and PS-SC, show consistent vertical gradients along the Lagrangian trajectory.

Lagrangian analysis with CALIPSO suggests that such extreme events have the potential to ascend air parcels up to the free troposphere, above 350 K potential temperature height. Once the air parcel is within the 350 K potential temperature height, it can continue to ascend with slow diabatic heating in this region. North of 32° N over the TP, the 350 K potential temperature line crosses the tropopause, as observed by CALIPSO. This may create an efficient pathway for the exchange of water vapour and other pollutants from the upper troposphere to the lower stratosphere through horizontal advection. In addition, deep convection, along with orographic lifting, also helps to reach air parcels near the tropopause region, as observed by E2. A previous study based on long-term satellite observations and Lagrangian trajectory analysis concluded that the overall effect of deep convection on lower stratospheric moistening is very small in the current scenario, but in a changing climate, it can be substantially high (Ueyama *et al.*, 2023).

4. Conclusions

Evaluation of three extreme events over the Third Pole region demonstrates that ICON-CLM simulations at km-scale capture precipitation more realistically than parametrised-scale simulations or reanalysis. Although ERA5 often exhibits similar or even higher CAPE values, it fails to capture extreme precipitation as effectively as km-scale simulations, particularly those with the shallow convection parametrisation (KM-SC). This highlights the benefit of explicitly resolving convection in future regional studies over the Third Pole region.

ICON-CLM km-scale simulations captured the spatial distribution of specific humidity at 100 hPa during extreme events E1 and E2, performing better than parametrized-scale simulations. Although ERA5 shows higher values at 100 hPa in E2, differences compared to satellite observations still remain. ICON-CLM simulations show an increase in specific humidity in the upper troposphere and at 100 hPa (Fig. 4) proportional to CAPE in the region (1600 J/kg, 1200 J/kg, and 400 J/kg for E1, E2, and E3, respectively), but this correspondence is not observed in ERA5. In addition, Figure 8 shows that along the CALIPSO path, the ICON-CLM simulations exhibit high specific humidities up to 100 hPa in the region of deep convection. In contrast, ERA5 depicts the lowest humidity in the convective region in E1 and E2 but higher values outside the event region. This indicates that a higher instability can increase the water vapour in the upper troposphere, as shown in km-scale simulations, but not necessarily in parametrised scale simulations and ERA5. Comparison of specific humidity from ICON-CLM and ERA5 with CALIPSO observations suggests that ERA5 exhibits a moisture bias in the upper troposphere, regardless of the strength of localised convection and may be influenced by the synoptic ASMA flow. This indicates that the enhancement of specific humidity during these events are primarily due to horizontal advection.

The analysis of water vapour flux during the events shows that vertical flux over the event region increases specific humidity by about 5 – 10% in the upper troposphere, but only by $\leq 1\%$ above the tropopause (100 hPa). Lagrangian trajectories show that only a few air parcels reach the upper troposphere (above 300 hPa) during the first few hours after initialization (Fig. 9). Once air parcels reach the upper troposphere, they tend to advect horizontally. The Lagrangian backward trajectories initiated above 100 hPa move westward before exiting the domain, resembling the circulation pattern of the ASMA. Back-trajectories complement forward trajectories, that deep convective events can stimulate a few air parcels to reach the upper troposphere. Then, they slowly move to

the lower stratosphere due to processes unrelated to the considered deep convective events. This result agrees with previous studies, which suggest that deep convection can uplift water vapour to 350 K and then in mid-latitude moisture advect horizontally (Hanumanthu *et al.*, 2020; Ploeger *et al.*, 2013; Singh *et al.*, 2020; Singh & Ahrens, 2025; Yan *et al.*, 2021).

This study focused on precipitation and snowfall based extremes, which were not necessarily the strongest deep convection events in the region. The ICON-CLM simulations' and ERA5's CAPE values also indicate that the strength of convection was moderate for these events. During these events, air parcels reached the upper troposphere, but not able to cross the tropopause. Our results indicate that km-scale simulations more realistically capture the influence of local atmospheric instability on upper tropospheric moistening and vertical transport, whereas parametrised scale models, including ERA5, are more strongly influenced by synoptic-scale advection. These findings motivate km-scale modelling in further studies targeting better understanding of extreme convective events and associated transport of air parcels in the Third Pole region.

Data availability

One set of simulations, KM-SC, is available at CPTP-CORDEX as contribution number 17 (http://rcg.gvc.gu.se/cordex_fps_cptp/); the other data set can be obtained by writing to Prof. Bodo Ahrens (Bodo.Ahrens@iau.uni-frankfurt.de).

Code and data availability

Visualisation codes can be accessed at <https://doi.org/10.5281/zenodo.14771288>.

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Authors' contributions

Prashant Singh: Conceptualization, methodology, software, data curation, formal analysis, investigation, visualization, writing—review and editing.

Bodo Ahrens: Conceptualization, resources, project administration, funding acquisition, review and editing. (email: Bodo.Ahrens@iau.uni-frankfurt.de).

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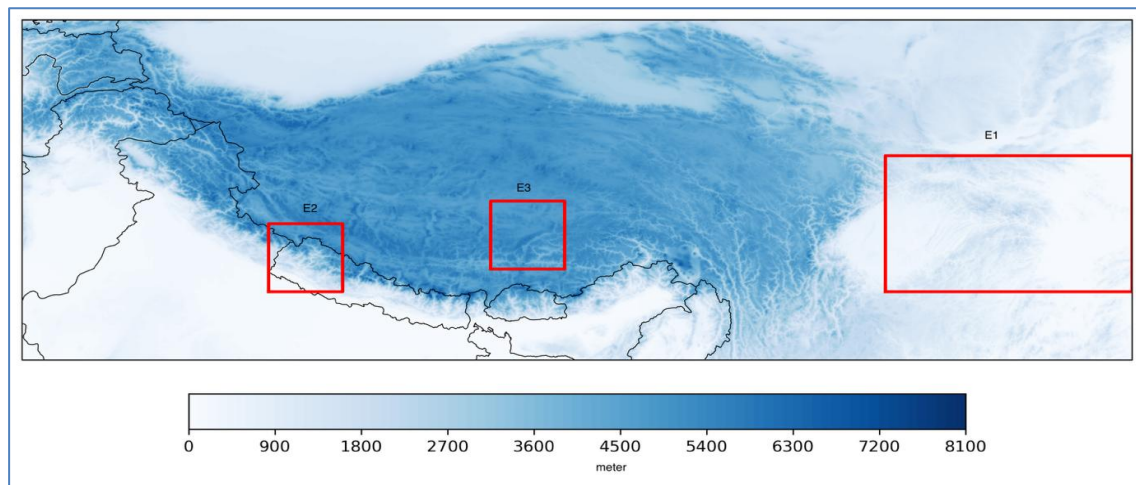
Supporting Figures

Fig. 1. Orography of the Third Pole domain with highlighted areas used for extreme event analysis at km-scale resolution (KM-SC)

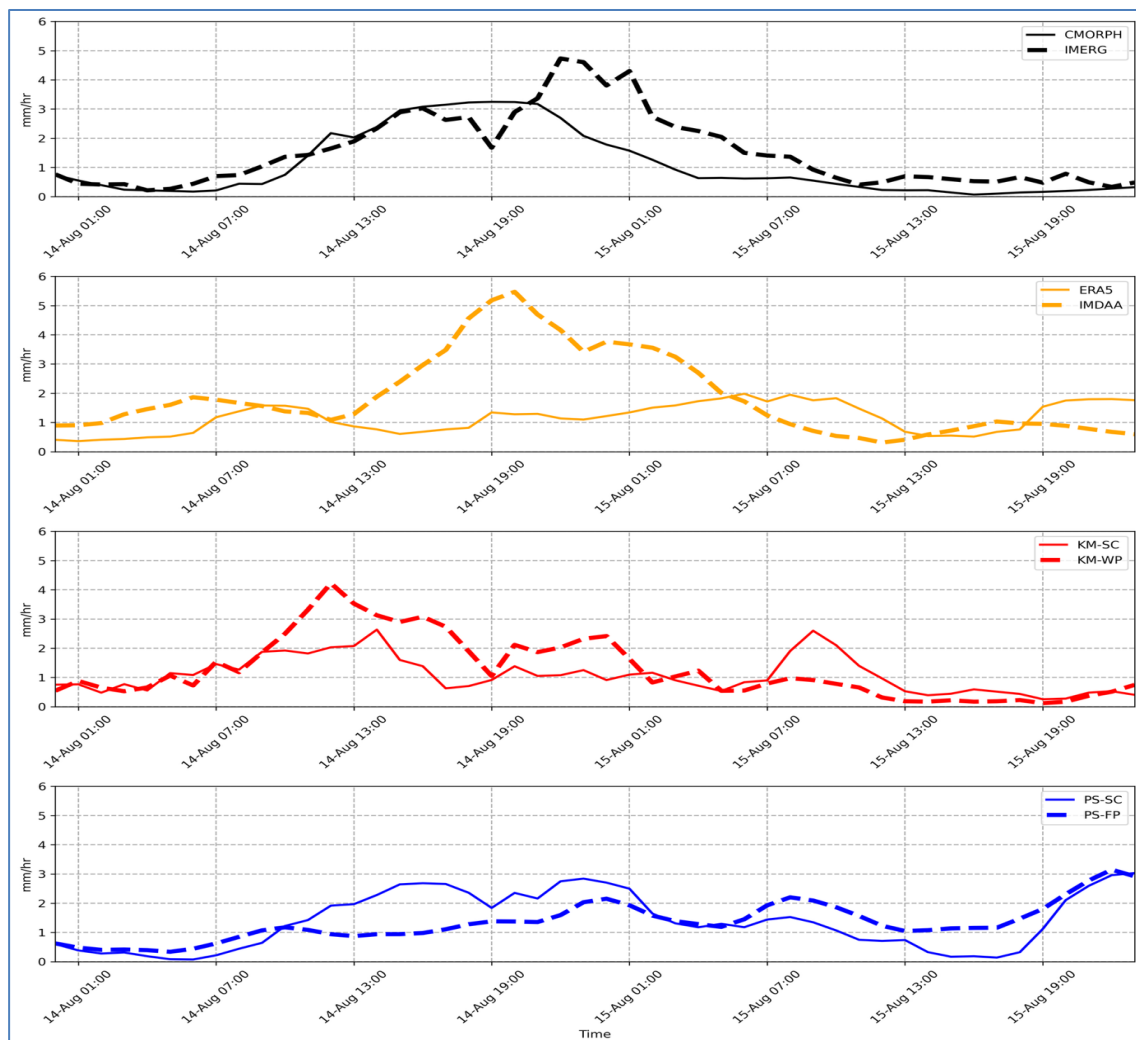


Fig. 2. Hourly precipitation area average over the second event (E2) from observations (black), reanalysis (orange), and ICON-CLM simulations at km-scale resolutions (red) and 13 km resolutions (blue)

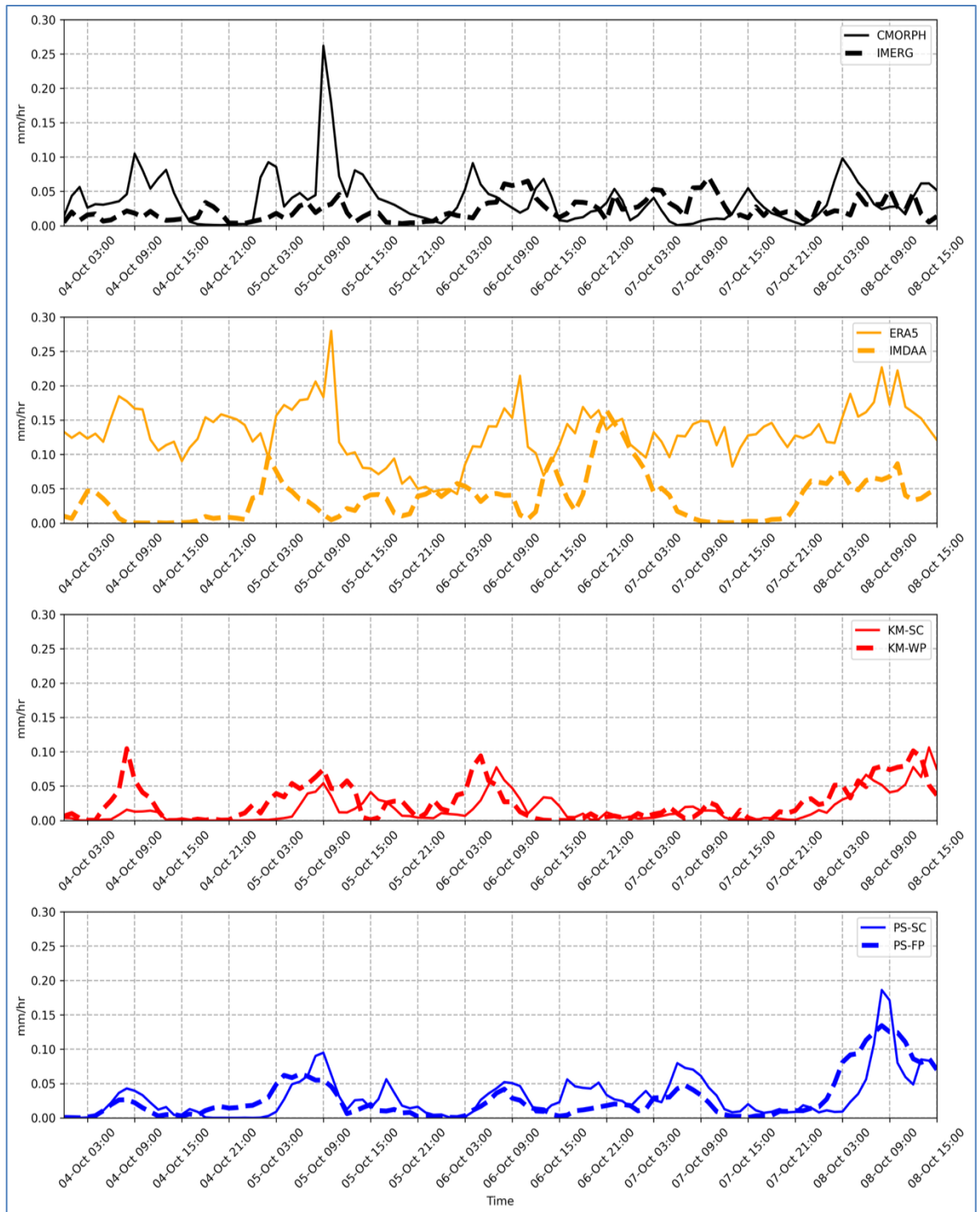


Fig. 3. Hourly precipitation area average over the third event (E3) from observations (black), reanalysis (orange), and ICON-CLM simulations at km-scale resolutions (red) and 13 km resolutions (blue)

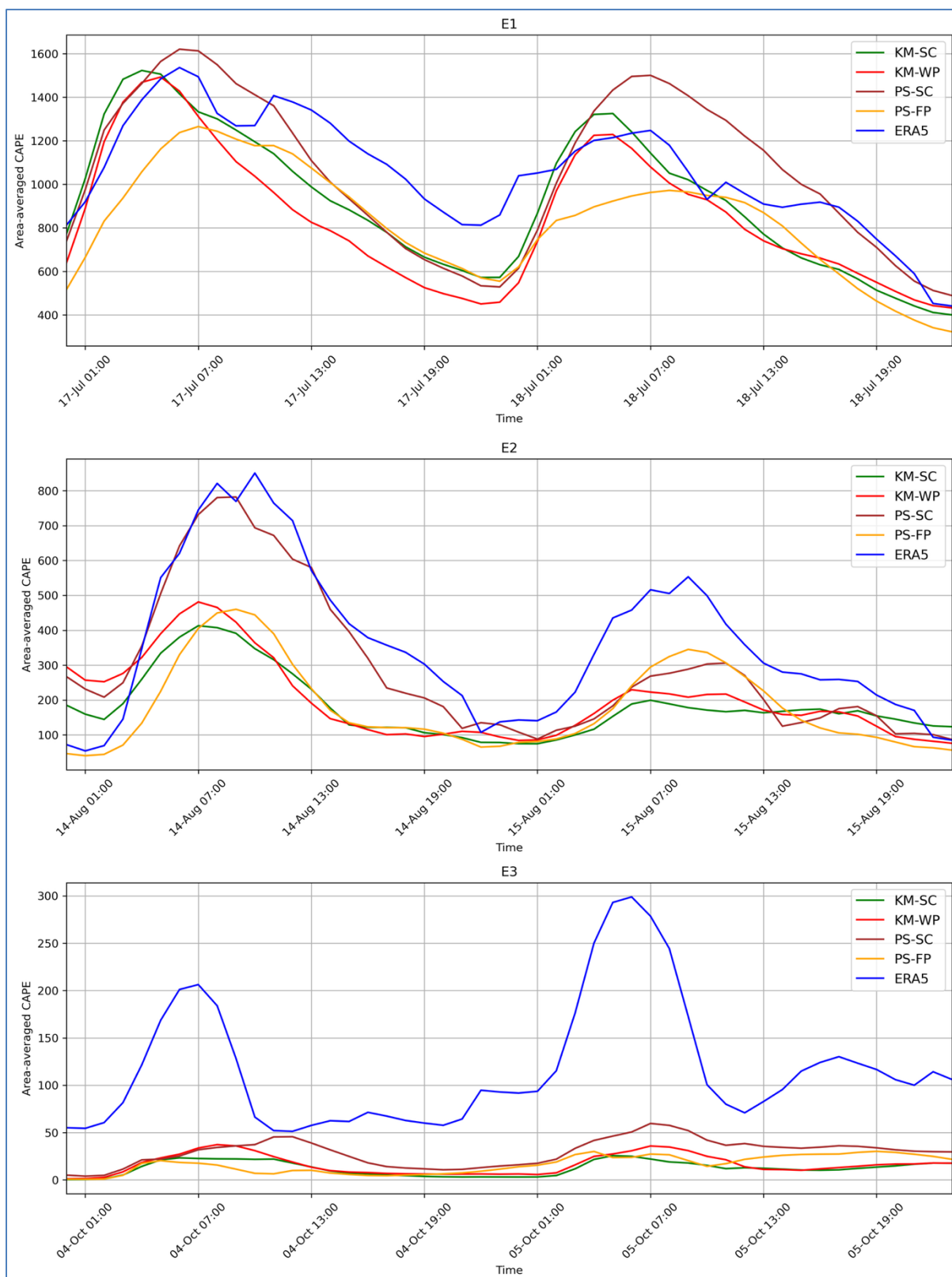


Fig. 4. Hourly CAPE area average over the events E1, E2, and E3 from ERA5 reanalysis (blue), and ICON-CLM simulations at 13 km resolutions and km-scale resolutions

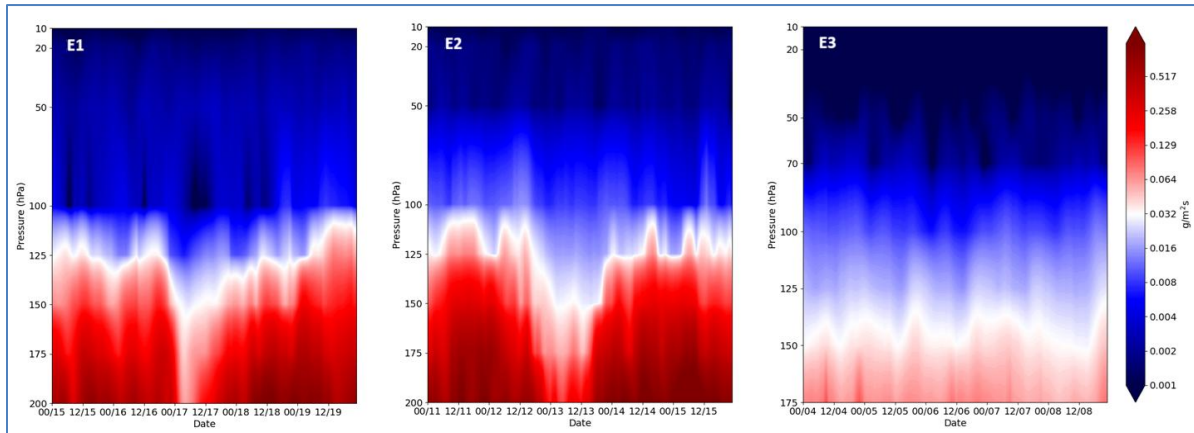


Fig. 5. Total water vapour flux for all three events averaged over the selected region of extreme events in ERA5

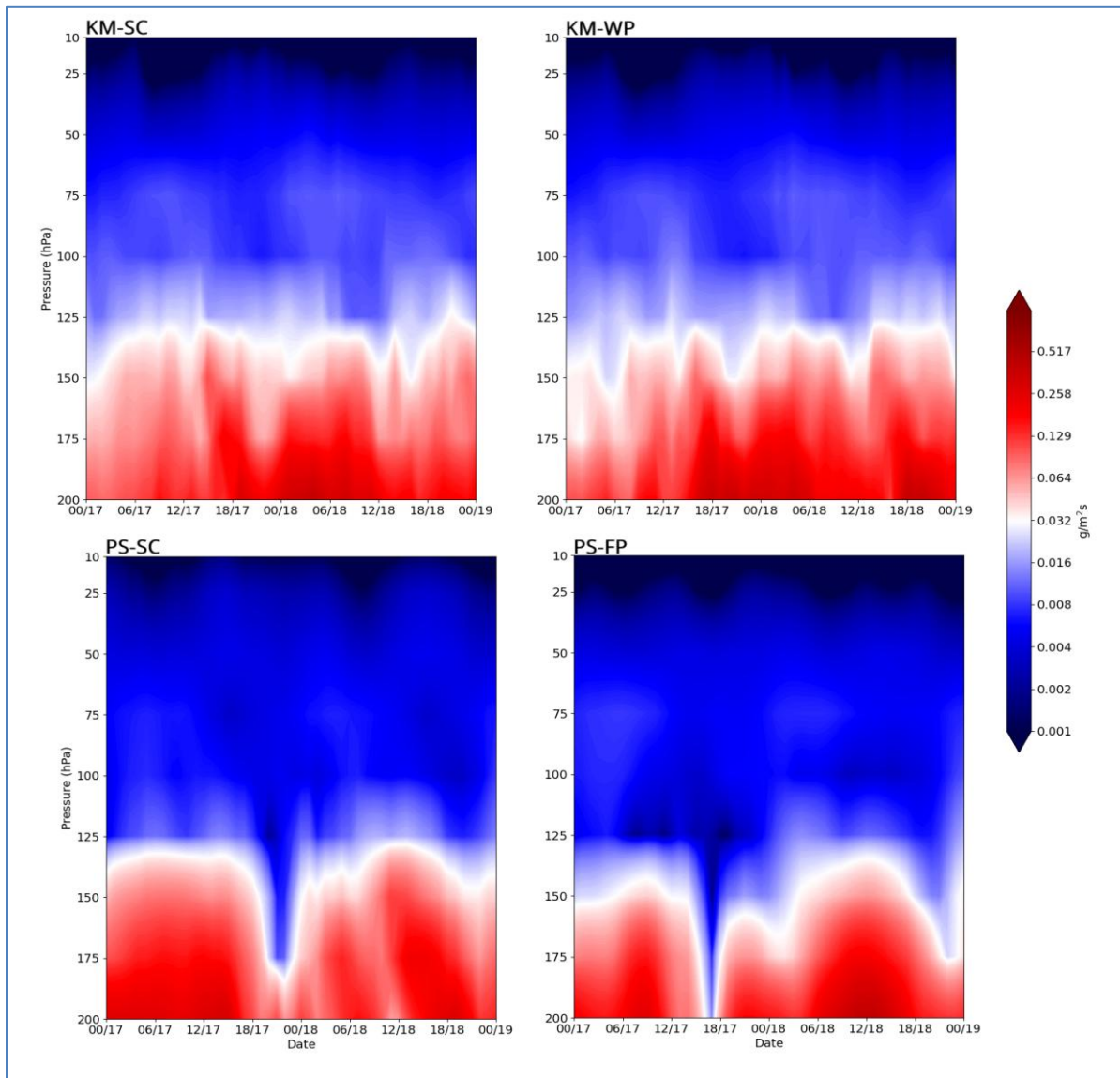


Fig. 6. Total water vapour flux averaged over the region of event E1 in KM-SC, KM-WP, PS-SC, and PS-FP, respectively