



A global assessment of warm-phase convective invigoration from aerosol-cloud interactions. Part 2: Observations

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Abstract. The hypothesis of warm-phase convective invigoration from aerosol-cloud interactions posits that aerosol pollution increases the cloud-droplet number concentration in the liquid layer of convective clouds, causing rising air parcels to consume supersaturation more readily. This leads to smaller supersaturations, larger buoyancies, and faster updrafts. We consider this mechanism to be substantial if it increases the updraft speed w by 25% or more when the cloud-droplet number concentration increases by a factor of five. A theoretical prerequisite for substantial invigoration is that supersaturation must exceed $\sim 0.9\%$ if $w = 3 \text{ m s}^{-1}$ and $\sim 5\%$ if $w = 7 \text{ m s}^{-1}$. This work searches for evidence of these high supersaturations in 12 globally distributed aircraft campaigns that collectively measure shallow and deep convection over ocean, coasts, and land. The quasi-steady approximation is applied to estimate supersaturation in liquid elements of convective updrafts with $\sim 100 \text{ m}$ horizontal resolution. Across the 13,342 total samples, 463 samples have supersaturation between 1% and 5%, seven samples have supersaturation between 5% and 9%, and no samples have supersaturation exceeding 9%. Cloud susceptibility to warm-phase invigoration is quantified as the fraction of updrafts with supersaturation exceeding the threshold for substantial invigoration. 62% of oceanic updrafts are susceptible when $1 \leq w < 2 \text{ m s}^{-1}$, but only 1% are susceptible when $w > 5 \text{ m s}^{-1}$. 14% of coastal and continental updrafts are susceptible when $1 \leq w < 2 \text{ m s}^{-1}$, but only 0.05% are susceptible when $w > 5 \text{ m s}^{-1}$. These results imply that precipitating shallow cumuli over the ocean are uniquely susceptible to warm-phase convective invigoration.

1 Introduction

Observations suggest that aerosol-cloud interactions may influence convective cloud dynamics. Convective cloud tops rise in conjunction with aerosol loading in some settings, and anomalous lightning flash rates often maximize over shipping lanes, cities, islands, and other localized aerosol sources (Orville et al., 2001; Andreae et al., 2004; Koren et al., 2010; Thornton et al., 2017; Wright et al., 2025). One hypothesis to explain these observations is called convective



invigoration. According to this idea, if cloud condensation nuclei become more abundant, then they perturb cloud microphysics in ways that cause updrafts to rise faster (Altartatz et al., 2014). Testing this hypothesis could contribute to the goal of understanding how aerosols influence convective mass flux, precipitation, and the radiative forcing of climate.

Three prevailing mechanisms have been proposed for convective invigoration. The first argues that aerosols inhibit precipitation from the collision and coalescence of cloud droplets in liquid clouds. These clouds thus lose more water through evaporation instead of precipitation. This moistens the environment and weakens the effects of entrainment, causing subsequent updrafts to retain their buoyancy over greater distances as they rise (Abbott and Cronin, 2021; Chua and Ming, 2020). The second proposed mechanism argues that aerosols reduce the collision and coalescence of liquid particles in the updrafts of deep convective clouds. Updrafts thus lift a greater mass of liquid particles above the freezing level and release more latent heat aloft when the particles freeze. This can cause the updraft to rise faster if the buoyancy gained from the additional latent heating is greater than the buoyancy lost from the additional condensate loading (Rosenfeld et al., 2008). These buoyancy anomalies typically offset one another quite closely, however (Grabowski and Morrison, 2016; Igel and van den Heever, 2021). Both of these mechanisms have only been demonstrated in idealized modeling frameworks, raising concerns over whether they generalize to the real atmosphere (Dagan et al., 2022; Peters et al., 2023).

The third proposed mechanism, and the focus of this study, is called warm-phase convective invigoration. According to this idea, updrafts in the liquid part of convective clouds can attain substantial supersaturation in non-polluted environments with sufficiently low aerosol concentration. If such clouds are exposed to aerosol pollution, then the resulting increase in cloud-droplet number concentration causes cloud parcels to consume supersaturation more readily. This lowers the supersaturation, causing convective air parcels to release latent heat earlier, to attain larger buoyancies, and ultimately to rise faster. The importance of this mechanism depends crucially on the supersaturation in non-polluted clouds because this sets an upper limit for how much aerosol pollution can change the supersaturation and buoyancy (Grabowski and Jarecka, 2015; Igel and van den Heever, 2021; Romps et al., 2023). Evidence of supersaturation that is large enough to support meaningful invigoration has so far come primarily from numerical simulations (Fan et al., 2018; Cotton and Walko, 2021; Fan et al., 2026).

Part 1 of this work derives an analytical model for warm-phase convective invigoration (Wall, 2026; hereafter Part 1). The model represents a rising thermal in the liquid part of a convective cloud. Warm-phase invigoration is predicted for a given change in the cloud-droplet number concentration with the meteorological environment held fixed, and invigoration is considered to be substantial if it causes the updraft speed to increase by 25% or more in response to a five-fold increase in the cloud-droplet number concentration. Two key predictions follow from this line of reasoning. First, for a given supersaturation, temperature, pressure, thermal size, and drag coefficient for the thermal, there exists an optimal updraft speed that experiences the largest amount of warm-phase invigoration. Second, substantial invigoration can occur only if the supersaturation exceeds a threshold that depends on the properties of the thermal. These predictions raise two questions: Under what circumstances are convective updrafts close to the optimal updraft speed, and how often do convective updrafts



65 have supersaturation that exceeds the theoretical threshold for substantial invigoration? This paper seeks to answer these questions.

From an observational perspective, the main challenge is to accurately quantify supersaturation in the liquid part of convective clouds (Varble et al., 2023, Part 1). The most accurate method at present is to measure the cloud-droplet size distribution and vertical wind speed with instrumented aircraft and infer the supersaturation using the quasi-steady approximation (Squires, 1952). This method has been applied to convective clouds over Australia (Warner, 1968), the Netherlands (Siebert and Shaw, 2017), India (Prabha et al., 2011; Fan et al., 2025), the Amazon (Romps et al., 2023, Romps, 2025a), the Philippines (Pan et al., 2026), and the interior and southern coast of the United States (Paluch and Knight, 1984; Politovich and Cooper, 1988; Patil et al., 2026). The supersaturation estimates from these studies are usually between 0.1% and 1%, but a handful of cases reach values on the order of 10% (Paluch and Knight, 1984; Prabha et al., 2011; Fan et al., 2025; Pan et al., 2026; Patil et al., 2026). The rare cases of high estimated supersaturation suggest that some clouds may be susceptible to substantial warm-phase invigoration.

Although these studies establish a baseline for the supersaturation in convective clouds, each study has focused on one location, and the majority of observations were obtained over land in updrafts slower than 10 m s^{-1} . For this reason, it has been challenging to examine a wide range of convective environments and identify the conditions that are most favorable for high supersaturation. We address this challenge by estimating supersaturation with observations from 12 globally distributed aircraft campaigns that collectively measure shallow and deep convection over land, coasts, and ocean. We then interpret the observations in the context of the analytical model from Part 1. This facilitates a global assessment of cloud susceptibility to warm-phase convective invigoration.

2 Observations

2.1 Aircraft Campaigns and Instrumentation

85 We analyze observations from 12 aircraft campaigns that penetrated the liquid part of convective clouds. The campaigns are outlined in Table 1. Three campaigns measured continental clouds. CuPIDO measured the transition from shallow to deep convection over a mountain in Arizona during the summer monsoon, and SPICULE and CloudGPS measured cumulus congestus over the plains of Colorado and Wyoming, respectively. Two other campaigns measured coastal clouds. COPE and ESCAPE measured deep convective clouds in sea-breeze convergence zones over the southwestern peninsula of the United Kingdom and the coast of Texas, respectively. The remaining seven campaigns measured oceanic clouds. CAMP²Ex measured small cumuli and cumulus congestus near the Philippines. ICE-T and SEAC⁴RS measured cumulus congestus and deep convective clouds near the United States Virgin Islands and over the Gulf of Mexico, respectively. RICO targeted precipitating cumuli with tops below the freezing level near Antigua. EUREC⁴A and CSET measured shallow trade-wind cumuli near Barbados and Hawai'i, respectively. Finally, SOCRATES targeted shallow cumuli in the cold sector of extratropical cyclones over the remote Southern Ocean. The campaigns over Arizona, Colorado,



and Wyoming are grouped in a “land” composite; the campaigns over Exeter and Texas are grouped into a “coast” composite; and the remaining seven campaigns are grouped into an “ocean” composite. Collectively these campaigns sample a wide range of meteorological environments and aerosol concentrations (Fig. 1).

In situ measurements are obtained from standard probes. Temperature T and ambient total air pressure p are measured by thermometers and pressure transducers, respectively. Vertical wind speed w is inferred from differential pressure measurements from either a gust probe or a combination of ports on the aircraft radome and pitot tubes on the fuselage. The size distribution of cloud droplets with diameter smaller than $\sim 50 \mu\text{m}$ is measured by a light-scattering probe, including either a Forward Scattering Spectrometer Probe-100 (FSSP), a Cloud Droplet Probe (CDP), a standalone Fast Cloud Droplet Probe (FCDP), or a Fast Cloud Droplet Probe contained within a Hawkeye Combination Cloud Particle Probe (HFCDP). Bin spacing for the light-scattering probes are listed in Table 2. The size distribution of cloud particles with diameter larger than $\sim 50 \mu\text{m}$ is measured by an imaging probe, including either a Two-Dimension Cloud Imaging Probe (2D-C), a Cloud Imaging Probe (CIP), or a Two-Dimensional Stereo spectrometer (2D-S). Finally, cloud water content is measured by a King or Nevzorov hot-wire probe. Among these data, the measurements of w and the size distribution of cloud droplets smaller than $\sim 50 \mu\text{m}$ are most important for estimating supersaturation. Measurement techniques, instrument comparisons, and accuracies for w and the cloud-droplet size distribution from light-scattering probes are synthesized by Lenschow (1986) and Baumgardner et al. (2017). Although accuracy depends on the instrument and atmospheric context, estimates of absolute accuracies range from 0.12 to 0.5 m s^{-1} for w ; 10% to 50% for the cloud-droplet number concentration, n ; and 10% to 30% for the mean cloud-droplet diameter, $\bar{D}$ (1- σ accuracy at 1-Hz sampling frequency) (Cooper et al., 2016; Baumgardner et al., 2017; Faber et al., 2018). All of the in-situ data are provided with sampling frequency between 1 and 25 Hz.

We also use measurements from radar onboard the aircraft. The campaigns over Arizona (CuPIDO), Wyoming (CloudGPS), Exeter (COPE), Antigua (RICO), and the United States Virgin Islands (ICE-T) use the Wyoming Cloud Radar (WCR) (Wang et al., 2012). The campaigns over Colorado (SPICULE), Hawai’i (CSET), and the Southern Ocean (SOCRATES) use the NSF NCAR High-performance Instrumented Airborne Platform for Environmental Research Cloud Radar (HCR) (NCAR/EOL HCR Team, 2014). The Texas campaign (ESCAPE) uses the National Research Council of Canada Airborne Radar (NAWX) (Wolde and Pazmany, 2005), the Barbados campaign (EUREC⁴A) uses the Radar System Airborne (RASTA) (Bony et al., 2022), and the Philippines campaign (CAMP²Ex) uses the Airborne Precipitation and Cloud Radar Third Generation (APR3) (Sadowy et al., 1997). All of the systems include a W-band radar that operates at a frequency of approximately 95 GHz. Vertically-oriented Doppler-velocity profiles are selected from the W-band radar. Most of the profiles are nadir looking, but zenith-looking profiles are considered in cases where nadir-looking profiles are not available. Doppler velocity is reported with a vertical resolution of 15 to 75 m, sampling frequency of 0.5 to 30 Hz, and a blind range between the aircraft and the first available data point of 75 to 390 m (Table A1). W-band radar is not available for the Gulf of Mexico campaign (SEAC⁴RS). The aircraft instrumentation is summarized in Table 3.



Radiosonde measurements are obtained from the University of Wyoming Atmospheric Science Radiosonde Archive (Table A2). Soundings are launched every 6 to 12 hours. For each campaign, we select the sounding station that is closest to the campaign headquarters and located at an elevation where the lifting condensation level is representative of the sampled clouds. Aircraft measurements are then matched with the closest sounding in time. Finally, the temperature and pressure of the lifting condensation level are obtained from each sounding.

Data are mapped to a common time array. In most cases, in-situ data are provided on a common time array with a resolution of 1 s, and no additional mapping is performed. Four campaigns did not provide all of the in-situ data in such a format, however. The campaigns over the Philippines (CAMP²Ex), Wyoming (CloudGPS), and Texas (ESCAPE) include periods in which at least one variable derived from in-situ measurements is reported with temporal resolution shorter than 1 s. In these cases, data are averaged over 1-s intervals that match the other instruments. The campaign over the Gulf of Mexico (SEAC⁴RS) includes w with 1-s resolution on a time array that is offset from the other in-situ instruments. In this case, w data are linearly interpolated in time to match the time array of the other instruments. Radar data with temporal resolution shorter than 1 s and longer than 1 s are averaged and linearly interpolated in time, respectively, to 1-s intervals that match the in-situ data. The final dataset has a common time array with resolution of 1 s. This corresponds to a horizontal resolution of ~ 100 m.

After mapping the data to a common time array, two adjustments are made to mitigate data biases. First, we remove the mean and linear trend of the in-situ measurements of w from each research flight. This corrects for an artificial offset and instrument drift that are visible in the raw time series of some flights from the Antigua and Arizona campaigns (RICO and CuPIDO, respectively). The mean and trend are computed without the first and last 10 s of data to avoid spurious values at the endpoints. Second, we synchronize the clocks of the light-scattering probe and the instrument that measures w to avoid bias in supersaturation estimates derived from these measurements (Romps, 2025a). Clock synchronization is checked by computing lag correlations between w and total cloud water content reported by the light-scattering probe. Lag correlations are computed over the data points in which w exceeds 0 m s^{-1} and cloud water content exceeds 0 g m^{-3} at all lags between -5 s and +5 s in increments of 1 s. Because fast ascent is expected to be collocated with large water content, the clock offset is estimated by the lag at which w and water content attain maximum correlation. Clock offsets relative to the w measurements are +1 s for the FSSP instrument in the Exeter campaign (COPE), -2 s for the FCDP instrument in the Texas campaign (ESCAPE), +1 s for the FCDP instrument in the Gulf of Mexico campaign (SEAC⁴RS). The timeseries of the light-scattering probes are shifted by these lags to synchronize the clocks. No adjustments are made in the remaining 12 campaigns and light-scattering probes because they attain maximum correlation at lag 0 s.

The database of aircraft observations includes 150 research flights that were performed by seven different research aircraft (Table 1). We include as many of the research flights as possible in the database, but some flights are excluded because they lack sufficient data. A total of 29 research flights are excluded from the campaigns over Barbados (EUREC⁴A), the Philippines (CAMP²Ex), Wyoming (CloudGPS), Arizona (CuPIDO), Texas (ESCAPE), Antigua (RICO), and the United States Virgin Islands (ICE-T) because they have missing radar or in-situ data, and a total of 19 research flights are excluded



from the SEAC⁴RS campaign because they did not target convective clouds over the Gulf of Mexico. Reasons for excluding flights are explained in Table A3.

165 Although the database includes a large set of airborne observations, it does not sample some convective conditions that favor high supersaturation. The observations do not sample intense continental convection due to aircraft safety concerns. They also do not sample deep convection over the remote tropical oceans with low aerosol concentrations, such as the Intertropical Convergence Zone. Finally, the number of research flights varies across the campaigns, giving more statistical weight to some locations than others. In particular, the Antigua campaign (RICO) provides the most research
170 flights in the ocean composite. Because this campaign targeted the warm-rain process, precipitating shallow cumuli may be overly represented in the ocean composite relative to a random sample. These sampling limitations should be kept in mind when interpreting the results.

Campaign	Acronym	Location	Time	Aircraft	Number of Research Flights Analyzed	Land, Coast, or Ocean Composite
Cumulus, Photogrammetric, In Situ, and Doppler Observations Experiment (Damiani, 2008)	CuPIDO	Santa Catalina Mountains, Arizona, USA	July and August 2006	University of Wyoming King Air	15	Land
Secondary Production of Ice in Cumulus Experiment (Lawson, 2023)	SPICULE	Broomfield, Colorado, USA	May and June 2021	NSF NCAR Gulfstream V	10	Land
CloudGPS (Parish and Leon, 2013)	CloudGPS	Laramie, Wyoming, USA	May 2008	University of Wyoming King Air	3	Land
Convective Precipitation Experiment (Leon	COPE	Exeter, UK	July and August 2013	University of Wyoming King Air	14	Coast



et al., 2016)						
Experiment of Sea Breeze Convection, Aerosols, Precipitation, and Environment (Kollias et al., 2025)	ESCAPE	Houston, Texas, USA	June 2022	NRC Convair 580	11	Coast
Studies of Emissions and Atmospheric Composition, Clouds, and Climate Coupling by Regional Surveys (Toon et al., 2016)	SEAC ⁴ RS	Gulf of Mexico (flights from other areas not included)	August and September 2013	NASA DC-8	4	Ocean
Ice in Clouds Experiment-Tropical (Heymsfield and Willis, 2014)	ICE-T	United States Virgin Islands	July 2011	NSF NCAR C-130	7	Ocean
Rain in Shallow Cumulus over the Ocean (Rauber et al., 2007)	RICO	Antigua	December 2004 and January 2005	University of Wyoming King Air	28	Ocean
Elucidating the role of cloud–circulation coupling in climate (Bony et al., 2022)	EUREC ⁴ A	Barbados	January and February 2020	SAFIRE ATR42	9	Ocean
Cloud System Evolution in the	CSET	Sacramento, California, USA	July and August 2015	NSF NCAR Gulfstream	16	Ocean



Trades (Albrecht et al., 2019)		to Kona, Hawai'i, USA		V		
Cloud, Aerosol and Monsoon Processes Philippines Experiment (Reid et al., 2023)	CAMP ² Ex	Luzon, Philippines	August, September, and October 2019	NASA P-3B	18	Ocean
Southern Ocean Cloud Radiation and Aerosol Transport Experimental Study (McFarquhar et al., 2021)	SOCRATES	Southern Ocean	January and February 2018	NSF NCAR Gulfstream V	15	Ocean

175 **Table 1: Aircraft campaigns considered in this study. NSF stands for National Science Foundation, NCAR stands for National Center for Atmospheric Research, NASA stands for National Aeronautics and Space Administration, and NRC stands for the National Research Council Canada.**

Instrument	Acronym	Diameter Bins for Cloud-Droplet Size Distribution
Forward Scattering Spectrometer Probe-100	FSSP	3- μm bins from 1.5 to 43.5 μm and one 4- μm bin from 43.5 to 47.5 μm for a total of 15 bins*
Cloud Droplet Probe	CDP	1- μm bins from 2 to 14 μm and 2- μm bins from 14 to 50 μm for a total of 30 bins
Fast Cloud Droplet Probe	FCDP**	1.5- μm bins from 0 to 6 μm , 2- μm bins from 6 to 18 μm , 3- μm bins from 18 to 42 μm , and 4- μm bins from 42 to 50 μm for a total of 20 bins
Hawkeye Fast Cloud Droplet Probe	HFCDP**	Same as FCDP

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Table 2: Light-scattering probes used to measure the cloud-droplet size distribution. *The FSSP in the United States Virgin Islands campaign (ICE-T) has 1.41- μm bins from 1.1 to 43.4 μm . **FCDP and HFCDP are the same type of probe, but the FCDP



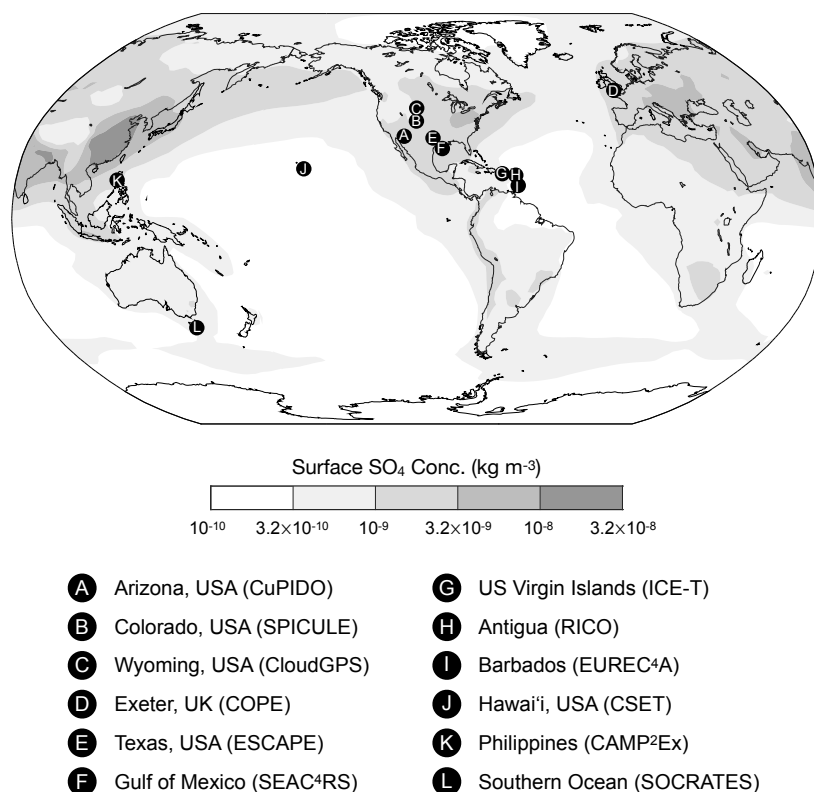
is a stand-alone probe and the HFCDP is contained within a Hawkeye Combination Cloud Particle Probe. They are given different names so that they can be distinguished when both are present on the same research aircraft.

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Campaign	In-situ w	Forward-scattering Probe	Imaging Probe	Hot-wire Probe	Radar
CuPIDO	gust probe	FSSP	2D-C	King	WCR
SPICULE	radome	CDP	2D-C	King	HCR
CloudGPS	gust probe	FSSP	2D-C	King	WCR
COPE	gust probe	FSSP and CDP	CIP	Nevzorov	WCR
ESCAPE	AIMMS-20	CDP and FCDP	-	Nevzorov	NAWX
SEAC ⁴ RS	MMS and AIMMS-20	FCDP	2D-S	-	-
ICE-T	radome	FSSP	2D-C	King	WCR
RICO	gust probe	FSSP	2D-C	King	WCR
EUREC ⁴ A	radome	CDP	2D-S	-	RASTA
CSET	radome	CDP	2D-C	King	HCR
CAMP ² Ex	TAMMS	FCDP and HFCDP	2D-S and 3V-CPI	-	APR3
SOCRATES	radome	CDP	2D-C	King	HCR

Table 3: Summary of aircraft instrumentation in the different campaigns. In the in-situ w column, “gust probe” and “radome” indicate that winds are obtained from differential pressure measurements from a gust probe on the nose boom and from radome ports and pitot tubes on the fuselage, respectively. “TAMMS” and “MMS” are the NASA Turbulent Air Motion Measurement System (Barrick et al., 1996) and Meteorological Measurement System (Scott et al., 1990), respectively. “AIMMS-20” is the Aircraft Integrated Meteorological Measurement System 20 Hz instrument. Acronyms for the other instruments are defined in Section 2.1. An entry of “-” indicates that the instrument is not available.

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195 **Figure 1: Locations of aircraft campaigns. Points show campaign locations, and contours show the dry mass concentration of sulfate aerosol at the surface averaged from 2003 through 2019 from MERRA-2 reanalysis.**

2.2 Quasi-Steady Supersaturation

A primary goal of this work is to quantify supersaturation, S . The most direct approach would be to compute S from
 200 measurements of the water-vapor partial pressure, T , and p . However, due to the limitations of current aircraft hygrometers, this approach quantifies relative humidity with 1- σ absolute accuracy of 2% to 10% outside of clouds (Zondlo et al., 2010). The presence of cloud droplets degrades the accuracy even further. It is difficult to draw strong conclusions about warm-phase convective invigoration with this degree of accuracy.

An alternative approach is to estimate S using the quasi-steady approximation. This approximation represents the
 205 supersaturation in a rising cloud parcel that would occur if the sink of water-vapor mixing ratio due to condensation is equal to the source of saturation water-vapor mixing ratio due to the ascending motion of the parcel. The condensational sink depends on S and the sum of the droplet diameters per unit volume, expressed as $n\bar{D}$, while the source depends on w . By relating the source and sink, it can be shown that



$$S = \frac{w}{n\bar{D}} \left[\frac{g}{2\pi R_a T} \left[\frac{1}{k_d} + \left(\frac{L}{R_v T} - 1 \right) \frac{L p_v^*}{R_v T^2 k_c} \right] \left(\frac{L R_a}{R_v T} - c_p \right) \left(c_p + \frac{R_a L^2 p_v^*}{R_v^2 T^2 p} \right)^{-1} \right], \quad (1)$$

where g is the gravitational acceleration, R_a is the specific gas constant of dry air, R_v is the specific gas constant of water vapor, k_d is the coefficient of diffusion for water vapor in air, k_c is the thermal conductivity of air, L is the latent heat of evaporation, p_v^* is the saturation water vapor pressure over liquid, and c_p is the heat capacity of dry air at constant pressure. Eq. 1 was derived by Romps (2025a), who modified the original derivation of Squires (1952). We compute p_v^* , $k_c(T)$, and $k_d(T, p)$ following theoretical derivations and fits to empirical data presented by Romps (2017; 2025a). With these relations, S is expressed as a function of w , n , $\bar{D}$, T , and p , all of which are measured in situ. The parameter F is a weak function of T and p that varies from 2.2 to 4.4 s m⁻³ over the range of conditions sampled.

The error of S derived from the quasi-steady approximation, considering only the measurement errors of the inputs, is given by $\left(\frac{\delta_S}{S}\right)^2 \approx \left(\frac{\delta_w}{w}\right)^2 + \left(\frac{\delta_n}{n}\right)^2 + \left(\frac{\delta_{\bar{D}}}{\bar{D}}\right)^2$, where δ is 1- σ absolute accuracy. Errors in T and p are neglected because they contribute very little to δ_S . Assuming a typical updraft speed of 5 m s⁻¹, $\frac{\delta_w}{w} \approx 0.1$. Modern CDP instruments have errors on the order of $\frac{\delta_n}{n} \approx 0.1$ and $\frac{\delta_{\bar{D}}}{\bar{D}} \approx 0.1$, while FSSP instruments have errors on the order of $\frac{\delta_n}{n} \approx 0.5$ and $\frac{\delta_{\bar{D}}}{\bar{D}} \approx 0.3$ (Baumgardner et al., 2017; Faber et al., 2018). Propagating these errors leads to an estimate of $\frac{\delta_S}{S}$ between 0.2 and 0.6. For $S = 1\%$, this corresponds to an absolute accuracy between 0.2% and 0.6%. This is an acceptable degree of accuracy for inferring cloud susceptibility to warm-phase invigoration.

To reliably estimate S with the quasi-steady approximation, care must be taken to identify instances in which the assumptions are valid (Romps, 2025a). Eq. 1 represents the supersaturation that a cloud parcel will adjust to if w , n , and $\bar{D}$ are fixed. This adjustment has a timescale τ_s that ranges from 0.1 to 10 s for typical liquid clouds (Korolev and Mazin, 2003). The quasi-steady approximation is thus valid if w , n , and $\bar{D}$ remain approximately constant over a time interval of τ_s prior to being measured by the aircraft. This condition is usually satisfied in the interior of convective clouds, even if droplets actively collide and coalesce inside the parcel (Romps et al., 2023). Situations that could violate this condition include activation of cloud condensation nuclei, rapid vertical acceleration of the updraft, or sedimentation of cloud droplets out of a parcel without replacement at the cloud edges (Korolev and Mazin, 2003). The quasi-steady approximation is also invalid in clear air, so aircraft samples that intersect cloud boundaries have invalid estimates of S .

Six filtering conditions are applied to select liquid-cloud updrafts that meet the conditions of validity for the quasi-steady approximation. All conditions are applied to the 1-Hz data unless stated otherwise. First, periods in which the aircraft is ascending, descending, or turning are excluded to avoid biases in the measurement of w (Atlas and Bretherton, 2023). Ascending and descending periods are identified by finding time segments that are at least one minute long in which the aircraft altitude increases or decreases by at least 1 m s⁻¹ in every 1-s interval within the time segment. Turns are identified similarly except that the aircraft roll angle is required to exceed 10°. Second, updrafts are selected by requiring that $w > 1$ m s⁻¹. Third, cloud interior points are selected by requiring that the cloud liquid mixing ratio q_l exceeds a threshold. q_l is



240 calculated from the size distribution of cloud droplets with diameter smaller than $\sim 50 \mu\text{m}$ measured by the light-scattering probe (FSSP, CDP, FCDP, or HFCDP). If high-rate data are available (10- or 25-Hz), then q_l sampled at the high rate is required to exceed $10^{-5} \text{ kg kg}^{-1}$ in every instance within a 1-s interval for that 1-s sample to be retained for analysis. If high-rate data are not available, then q_l sampled at 1-Hz is required to exceed $10^{-5} \text{ kg kg}^{-1}$ at time $t - 1$, t , and $t + 1$ for the sample at time t to be retained for analysis. This condition is designed to exclude samples at the lateral edges of clouds.

245 Fourth, liquid cloud elements are selected by requiring that $T > 0^\circ\text{C}$ or $\frac{\text{LWC}}{\text{TWC}} > 0.99$, where LWC and TWC are the liquid water content and total water content in units of g m^{-3} , respectively. Cloud water content is taken from hot-wire probes when available. If a Nevzorov probe is available, then measurements of LWC and TWC are taken from that probe. If a King probe is available, the LWC is taken from that probe and cloud ice water content is taken from a particle imaging probe. If no hot-wire probe is available, then samples are retained if $T > 0^\circ\text{C}$, thereby excluding supercooled liquid cloud. Fifth, samples

250 above the cloud base are selected by requiring that the altitude is at least a distance of $\Delta z = 250 \text{ m}$ above the lifting condensation level. This distance is an order of magnitude larger than the typical thickness of the layer in which cloud condensation nuclei activate at the cloud base (Pinsky and Khain, 2002). The condition is implemented by requiring that

$$p < p_{LCL} \exp\left(\frac{-g\Delta z}{R_d T_{LCL}}\right),$$

where p_{LCL} and T_{LCL} are the pressure and temperature of the lifting condensation level, respectively.

255 The sixth and final filtering condition requires the updraft to accelerate sufficiently slowly near the aircraft. Considering w fluctuations alone, the quasi-steady approximation is valid if $\left|\frac{dw}{dz}\right| \ll \frac{1}{\tau_s}$ in a reference frame following the air parcel (Korolev and Mazin, 2003). $\left|\frac{dw}{dz}\right|$ is approximated by $\left|\frac{dv_d}{dz}\right|$, where v_d is the radar Doppler velocity. By estimating $\left|\frac{dw}{dz}\right|$ this way, we are ignoring differences in the hydrometeor fall speed between adjacent radar bins, a distance of $\sim 30 \text{ m}$ (Table A1). τ_s is calculated with the in-situ data following Eq. 18 of Korolev and Mazin (2003). Finally, $\left|\frac{dw}{dz}\right|$ is required to be at

260 least one order of magnitude smaller than $\frac{1}{\tau_s}$ over every pair of adjacent radar bins in a vertical interval starting at the closest bin outside the radar blind zone and extending outward over a distance of $\max(3L_s, 100 \text{ m})$, where $L_s = w\tau_s$ is the supersaturation adjustment length-scale. A minimum distance of 100 m is applied to ensure that some radar bins are considered when L_s is shorter than the vertical resolution of the radar. We note that this method invokes the assumption of frozen turbulence: The profile of dw/dz at the instant an air parcel is measured is assumed to be representative of the recent

265 time history of the parcel.

Samples that satisfy all six conditions are retained for analysis. For these samples, the full cloud-droplet size distribution is computed by merging the bins of the light-scattering probe for $D \leq D_{max}$ and the bins of the imaging probe for $D > D_{max}$, where D is the droplet diameter and $D_{max} \approx 50 \mu\text{m}$ is the largest diameter measured by the light-scattering probe (Table 2). The lower bound of the size distribution measured by the imaging probe is truncated at D_{max} assuming that



270 the diameters of individual droplets are uniformly distributed within the bin containing D_{max} . $n\bar{D}$ is then computed from the merged distribution, and S is computed from Eq. 1.

Three exceptions are made due to missing data. First, the filtering condition that constrains dw/dz near the aircraft is not applied to the Gulf of Mexico campaign (SEAC⁴RS) because W-band radar is not available. Second, the filtering condition of $\frac{LWC}{TWC} > 0.99$ when $T < 0$ °C is not applied in the Philippines campaign (CAMP²Ex) because a hot-wire probe is
275 not available. Instead, we identify supercooled liquid by requiring that all particles are spherical based on visual inspection of images from the 3V-CPI. Third, the cloud-droplet size distribution is computed for droplets with $D \leq 50$ μm for the Texas campaign (ESCAPE) because a quality-controlled scientific dataset for the imaging probe was not available at the time of study. Excluding droplets with $D > 50$ μm leads to a small overestimation of S that is usually less than 0.01% (Appendix B).

280 The reproducibility of the S estimates can be assessed in cases where separate instruments measured the inputs of Eq. 1 at the same place and time. The campaigns over Exeter (COPE), Texas (ESCAPE), and the Philippines (CAMP²Ex) measured the cloud-droplet size distribution with two separate light-scattering probes, and the campaign over the Gulf of Mexico (SEAC⁴RS) measured w , T , and p with two separate instruments (Table 3). We compute S using both sources of input and select the samples in which both cases satisfy the filtering conditions for S . The fraction of samples in which the
285 coincident S estimates agree to within a factor of two ranges from 0.82 to 0.96 (Fig. 2). Thus, as a rule of thumb, each individual estimate of S at 1 Hz sampling frequency is likely accurate to within a factor of two, including random and systematic errors.

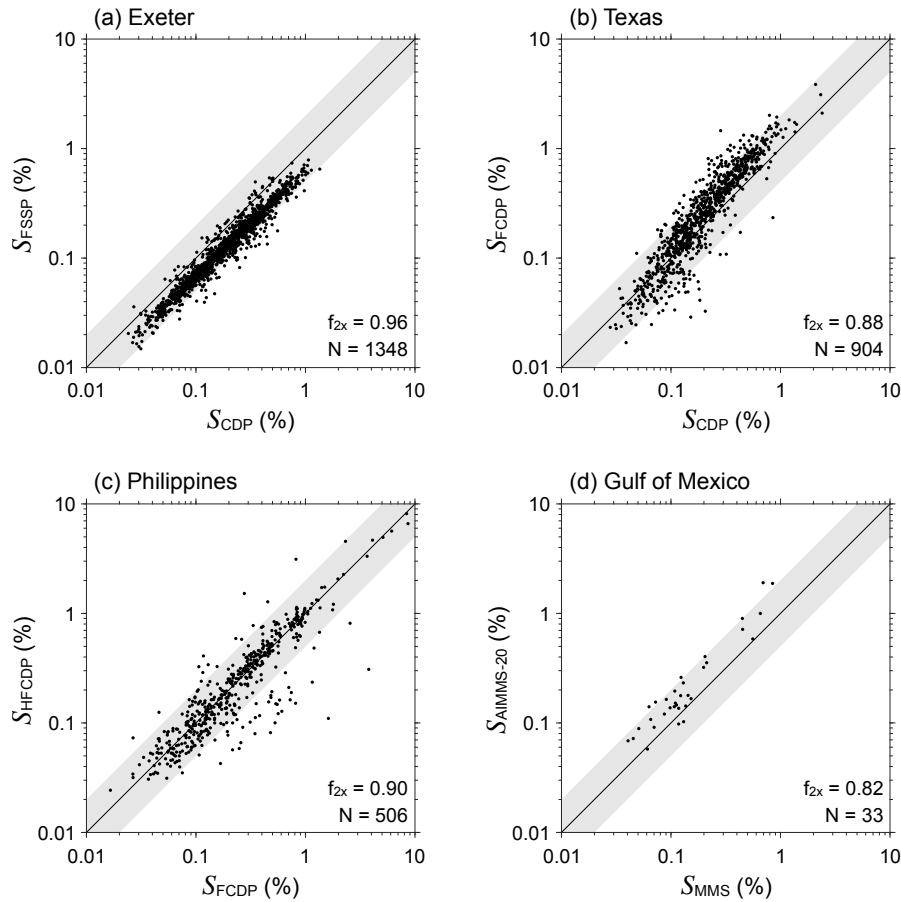


Figure 2: Reproducibility of S estimates. (a-c) Scatterplots of coincident S estimates computed with different light-scattering probes obtained in the Exeter, Texas, and Philippines campaigns (COPE, ESCAPE, and CAMP²Ex, respectively). The subscripts for S indicate the light-scattering probe (Table 2). Coincident S estimates that lie within the gray shading agree to within a factor of two. The fraction of samples that agree to within a factor of two (f_{2x}) and the sample size (N) are printed in the bottom right corner. (d) Similar to (a-c), but showing coincident S estimates computed with different meteorological instruments from the Gulf of Mexico campaign (SEAC⁴RS). The subscripts for S indicate the instruments that measured T , p , and w (Table 3).

3 Results

3.1 Supersaturation

We next examine the statistics of S in liquid elements of convective clouds. The values can be interpreted with the help of the analytical model from Part 1. This model considers a rising thermal in the liquid part of a convective cloud with radius r , vertical wind speed w , supersaturation S , and cloud-droplet number concentration n . Vertical perturbation pressure-



gradient forces acting on the thermal are represented by a drag law with drag coefficient c_d . The model predicts the increase in updraft speed that arises from warm-phase invigoration if the cloud-droplet number concentration increases from n to $\hat{n}$ while the meteorological environment remains fixed. Invigoration is considered substantial if it causes the updraft speed to increase by 25% or more when the cloud-droplet number concentration increases by a factor of five (i.e. $\frac{\hat{n}}{n} = 5$). This line of reasoning implies that substantial invigoration can occur only if S exceeds a threshold S_{min} that is given by

$$S_{min} = \frac{(\Lambda^2 + 2\Lambda)w^2 c_d}{(1 - \Lambda\beta - \beta)ra} \quad (2)$$

where $\beta = \left(\frac{\hat{n}}{n}\right)^{-2/3}$, a is a function of T and p , and $\Lambda = 0.25$ is the minimum relative increase in the updraft speed that constitutes substantial invigoration (i.e. 25%). $\frac{\hat{n}}{n} = 5$ is chosen to represent the range of typical variability observed in the cloud-droplet number concentration, while $\Lambda = 0.25$ is chosen arbitrarily (Boucher and Lohmann, 1995). If ra/c_d takes on a typical value of $1000 \text{ m}^2 \text{ s}^{-2}$, then an updraft with $w = 3 \text{ m s}^{-1}$ typical of shallow convection can experience substantial invigoration if S exceeds $S_{min} \approx 0.9\%$. A faster updraft with $w = 7 \text{ m s}^{-1}$ typical of oceanic deep convection can experience substantial invigoration if S exceeds $S_{min} \approx 5\%$. An even faster updraft with $w = 15 \text{ m s}^{-1}$ typical of continental deep convection can experience substantial invigoration if S exceeds $S_{min} \approx 22\%$. A key goal is to determine how often S exceeds these theoretical thresholds.

We first consider the distribution of S from all of the samples combined (Fig. 3). Instances in which S is estimated from two separate instruments are treated as two independent samples. The estimates of S are grouped into land, coast, and ocean composites that contain 2178, 5306, and 5858 samples, respectively. Land clouds have a relatively narrow range of S , including a median of 0.19% and a 99th percentile of 0.56%. Only one out of 2178 total samples has $S > 1\%$. Coastal clouds have a distribution with a similar median but a longer upper tail. In this case the median of S is 0.18%, the 99th percentile is 1.19%, and a fraction of 0.02 of the samples have $S > 1\%$. Oceanic clouds have an even greater median and range. In this case the median of S is 0.46%, the 99th percentile is 1.91%, and a fraction of 0.08 of the samples have $S > 1\%$. Across the 13,342 samples from all campaigns, 463 samples have $1\% \leq S < 5\%$, 7 samples have $5\% \leq S < 9\%$, and no samples have $S \geq 9\%$. Supersaturations that exceed the threshold of $S_{min} \approx 0.9\%$ for typical shallow convective updrafts are observed much more often than supersaturations that exceed the thresholds of $S_{min} \approx 5\%$ and $S_{min} \approx 22\%$ for typical oceanic and continental deep convection.

The observations can be stratified by the parameters in Eq. 1 to examine the factors that control S . The first such factor is w . Faster ascent cools air parcels more rapidly by expansion, leading to a greater source of supersaturation and larger values of S according to the quasi-steady approximation. The observations of w span 1 to 22 m s^{-1} . Across this range, w is positively correlated with S in every campaign except the Gulf of Mexico (SEAC⁴RS) (Fig. 4). However, a linear regression between w and S across all samples explains only 8% of the variance of S . In other words, faster ascent favors higher S , but w alone does not constrain S very tightly.



Another controlling factor for supersaturation is the sum of the droplet diameters per unit volume, expressed as $n\bar{D}$. This term is called the “diameter concentration” by Romps et al. (2023), and it is equivalent to twice the “integral radius” defined by Politovich and Cooper (1988). The condensation rate on a single cloud droplet in a convective updraft is proportional to S times the droplet diameter, and the condensation rate for a population of droplets is proportional to $Sn\bar{D}$ (Romps, 2025b). Parcels with relatively large $n\bar{D}$ thus consume supersaturation more readily and favor small S . The observed values of $n\bar{D}$ span 0.03 to 31.4 mm cm⁻³. The median is largest over land (5.8 mm cm⁻³), followed by coast (5.7 mm cm⁻³), and then ocean (3.3 mm cm⁻³). The parameter $(n\bar{D})^{-1}$ is positively correlated with S in every campaign, and a linear regression of $(n\bar{D})^{-1}$ against S across all of the samples explains 48% of the variance of S (Fig. 5). Furthermore, when $S > 1\%$, $n\bar{D}$ is usually less than 2 mm cm⁻³. These low values of $n\bar{D}$ occur much more frequently over ocean than over coast and land. This explains why the majority of samples with $S > 1\%$ occur over ocean (Fig. 3).

Variations of w and $n\bar{D}$ are expected to control the vertical structure of S inside the liquid layer of convective clouds. Rising air parcels experience a time tendency of $n\bar{D}$ that depends on the dominant droplet-growth mechanism. Condensational growth increases $n\bar{D}$, while collision and coalescence decreases it. The theoretical droplet coalescence rate is approximately proportional to the fifth power of the mean volume radius of the droplets (Freud and Rosenfeld, 2012). This strong dependence implies that collision and coalescence is typically negligible at the bottom of convective clouds and rapidly overtakes condensation as the dominant growth mechanism once the cloud droplets grow to a threshold size. It also implies that the condensational sink of S rapidly shrinks once collision and coalescence dominates, at which point S may dramatically increase. A convective cloud that is deep enough for the droplets to reach the collision-coalescence threshold is expected to have relatively small S near the cloud base and larger S at the top of the liquid layer. This top-heavy structure of S may be enhanced by the fact that w typically increases with height.

To test this idea, the observations of S are composited by the distance above cloud base. Data from each campaign are sorted into bins of $|p - p_{LCL}|$ with a width of 100 hPa, where p_{LCL} is the pressure of the lifting condensation level. Compositing the data this way reveals that the median S increases with height above the cloud base in the ocean and coast campaigns that sampled multiple bins (Fig. 6). Furthermore, S follows a concave vertical profile that includes nearly constant values in the lowest two bins of $|p - p_{LCL}|$ followed by progressively increasing values above. In contrast, S follows a nearly constant vertical profile in the land campaigns, except for a slight increase with height between the two highest bins in Arizona (CuPIDO). Oceanic and coastal clouds often reach the collision-coalescence regime below the freezing level, while continental clouds reach this level at a greater distance above the cloud base, if they reach it at all (Freud and Rosenfeld, 2012). Thus, the vertical structure of S is consistent with the expectation that S maximizes at upper levels in the liquid layer of convective clouds where droplets grow by collision and coalescence.

The importance of collision and coalescence can be further examined by compositing by cloud-base temperature. As the cloud base becomes warmer, the layer between cloud base and the freezing level deepens, and rising parcels have more time to reach the collision-coalescence regime before freezing occurs. The greatest supersaturations occur where



collision and coalescence is afforded the most time to reduce $n\bar{D}$ well below its value near cloud base. This is expected to occur most frequently near the freezing level in clouds with warm bases.

To examine whether cloud-base temperature affects the likelihood of high S , we sort the observations into warm-base and cold-base composites that are defined by whether the lifting condensation level is warmer or colder than 18 °C. This threshold is chosen so that the sample size is comparable in the two cases ($N = 7582$ and $N = 5760$, respectively). The samples from each case are then composited by T and w , and the probability of observing $S > 1\%$ is computed (Fig. 7). For most T - w combinations, $S > 1\%$ occurs much more frequently in warm-base clouds than in cold-base clouds. The probability of observing $S > 1\%$ maximizes at values around 0.36 near the freezing level in warm-base clouds with $w > 10$ m s⁻¹. In contrast, the probability of observing $S > 1\%$ in cold-base clouds is 0.04 or less in all T - w bins. This result shows that S regularly reaches values of ~1% near the freezing level when the updraft speed exceeds 10 m s⁻¹ and the liquid layer is deep enough for substantial collision and coalescence to develop.

The seven largest values of S are between 5% and 9%. These values exceed the theoretical threshold of $S_{min} \approx 5\%$ that is required for substantial invigoration of typical deep convective updrafts over ocean. For this reason, we examine these samples in more detail to determine their credibility. The samples were observed in four penetrations of cumulus congestus near the Philippines (CAMP²Ex). All were measured at temperatures colder than 0 °C, where the largest supersaturations are expected (Fig. 8). Furthermore, they were all determined independently with two light-scattering probes, and the two estimates agree to within a factor of 1.3 in every case. This suggests that the measurements of $n\bar{D}$ are credible and reproducible. The credibility of the S estimates thus depends on whether the quasi-steady approximation is valid in these instances.

The validity of the quasi-steady approximation can be evaluated based on the measurement position within the cloud. The first instance of $S > 5\%$ was observed in the horizontal midpoint of an updraft core on September 7, 2019 at 02:53:36 UTC (Fig. 8a). This observation appears to be located at the very top of a cumulus congestus cloud in forward-facing video footage from the front of the aircraft. Cloud-particle images taken at 400-Hz by the 3V-CPI indicate that no cloud droplets were present for 0.67 s within the 1-s interval over which S is estimated. Thus, the 1-s interval for this observation may have sampled both cloudy and cloud-free air near the cloud top. Because the quasi-steady approximation is not valid in cloud-free air, the estimate of S in this instance is probably not reliable. The second instance of $S > 5\%$ was observed at the horizontal edge of an updraft core on September 17, 2019 at 02:53:23 UTC (Fig. 8b). Like the previous case, this instance appears to be located at the top of a congestus cloud in forward-facing video, and no cloud particles are visible in 3V-CPI imagery for 0.69 s within the 1-s interval over which S is estimated. This estimate of S is thus unreliable as well. The third instance of $S > 5\%$ occurred in a weak updraft with maximum vertical wind speed of 2.0 m s⁻¹ that appears as a wispy, thin cloud filament in forward-facing video (Fig. 8c). This case appears to be a decaying convective element to the side of a prominent congestus cloud. The final instance of $S > 5\%$ occurred in the middle of an isolated congestus cloud on September 29, 2019 at 07:32:32 (Fig. 8d). The sampled air is located in the interior of the cloud well below the cloud top in



forward-facing video. However, it represents a single ~100 m segment between two updrafts, and it is not representative of
the surrounding updraft cores, in which S is at least three times smaller. In all of these cases, S exceeds 5% in one ~100 m
segment, but not over the entire updraft. Thus, we find no reliable evidence that S exceeds 5% over the majority of a rising
thermal in the clouds sampled in this work. An upper bound of $S \leq 5\%$ is a conservative limit for horizontal scales
comparable to the size of cloud updrafts.

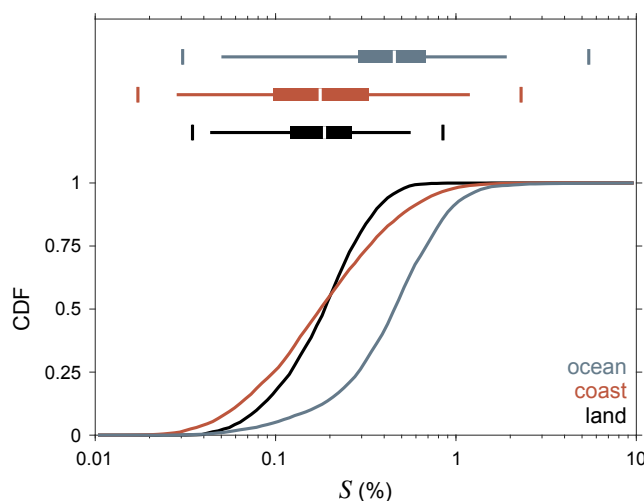


Figure 3: S statistics from all campaigns. Box plots on the top show the 1, 25, 50, 75, and 99 percentiles, and the outer vertical lines show the 0.1 and 99.9 percentiles. Lines on the bottom show the cumulative distribution function (CDF). Statistics are shown for all ocean campaigns, all coast campaigns, and all land campaigns (Table 1). The sample size for ocean, coast, and land campaigns is 5858, 5306, and 2178, respectively.

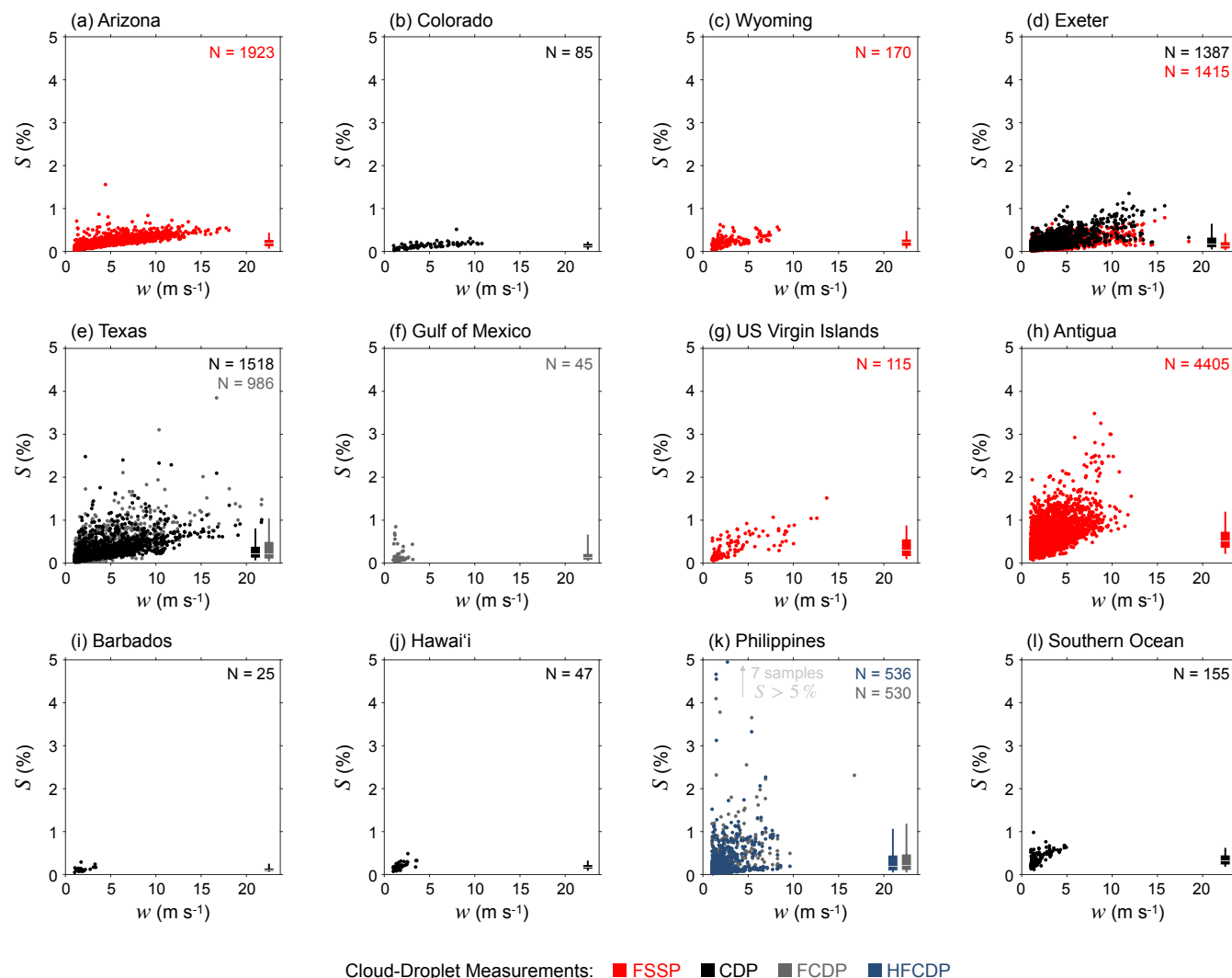


Figure 4: S as a function of w . Each panel shows data from one campaign. Box plots in the bottom right corner show the 5, 25, 50, 75, and 95 percentiles of S computed from all samples. Points are colored according to the light-scattering probe that measured the cloud-droplet size distribution. The sample size N is printed in the top right corner. Note that the Philippines campaign (CAMP²Ex) has seven samples with $S > 5\%$ that are not visible on the figure (k).

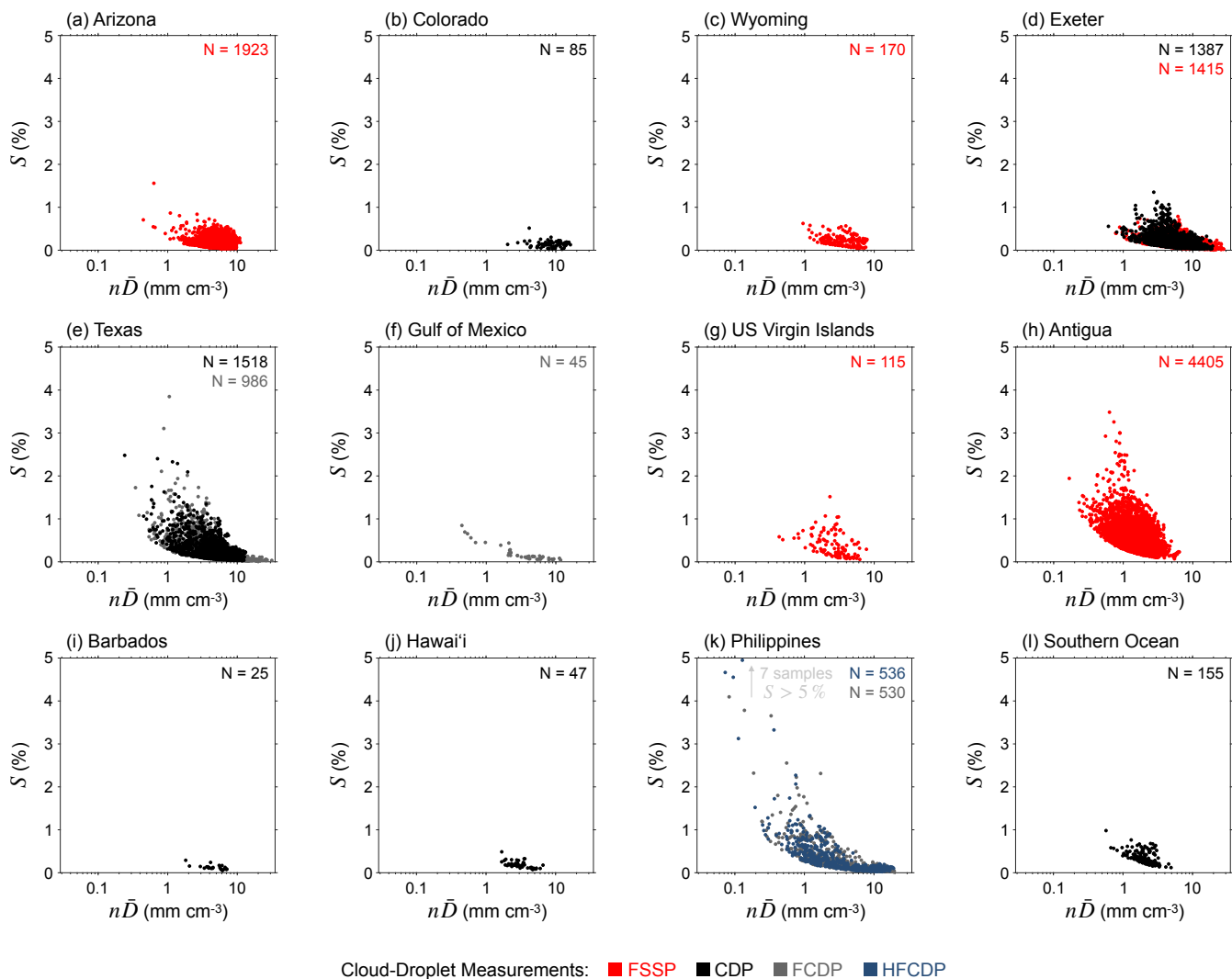


Figure 5: S as a function of $n\bar{D}$, where n is the cloud-droplet number concentration and $\bar{D}$ is the mean cloud-droplet diameter. Points are colored according to the probe that measured the cloud-droplet size distribution (see legend). Note that the Philippines campaign (CAMP²Ex) has seven samples with $S > 5\%$ that are not visible on the figure (k).

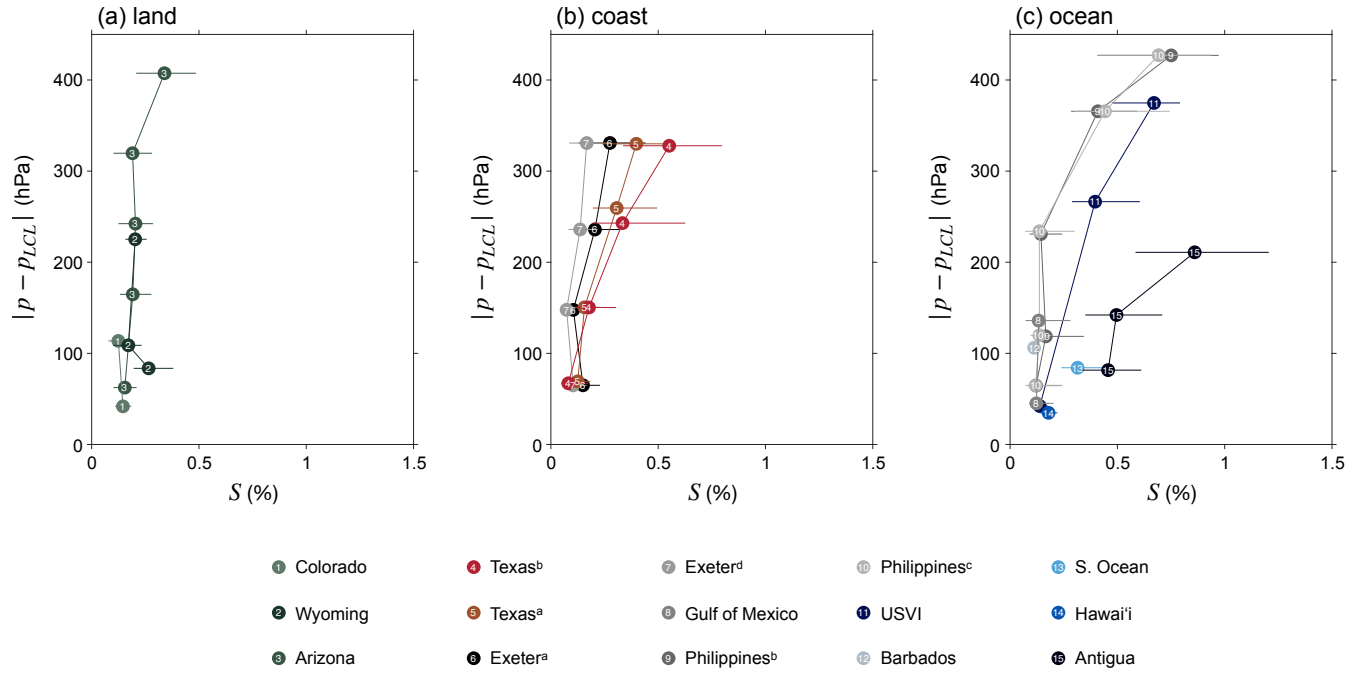


Figure 6: S as a function of the distance above cloud base. (a) S from land campaigns plotted as a function of $|p - p_{LCL}|$, where p_{LCL} is the pressure of the lifting condensation level. Data are binned by $|p - p_{LCL}|$ according to the values labelled on the vertical axis. Circles and horizontal lines show the median and interquartile range of S within a given pressure bin. (b-c) Similar to (a), but showing the coast and ocean campaigns, respectively. Superscripts a, b, c, and d indicate that S was computed using the cloud-droplet size distribution measured by the CDP, FCDP, HFCDP, and FSSP probes, respectively.

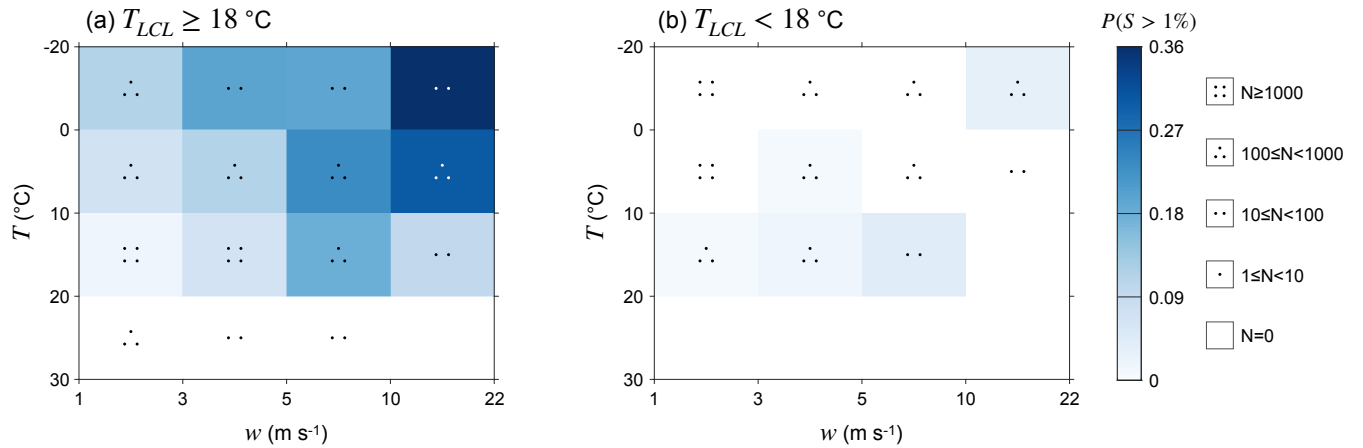


Figure 7: Probability of observing $S > 1\%$ as a function of T and w for (a) warm-base clouds and (b) cold-base clouds. Warm- and cold-base clouds are defined by $T_{LCL} \geq 18^\circ\text{C}$ and $T_{LCL} < 18^\circ\text{C}$, respectively, where T_{LCL} is the temperature of the lifting



condensation level. Dots indicate the number of observations in each bin (N) according to the legend. Data from all campaigns are included. The sample size is 7582 for warm-base clouds and 5760 for cold-base clouds.

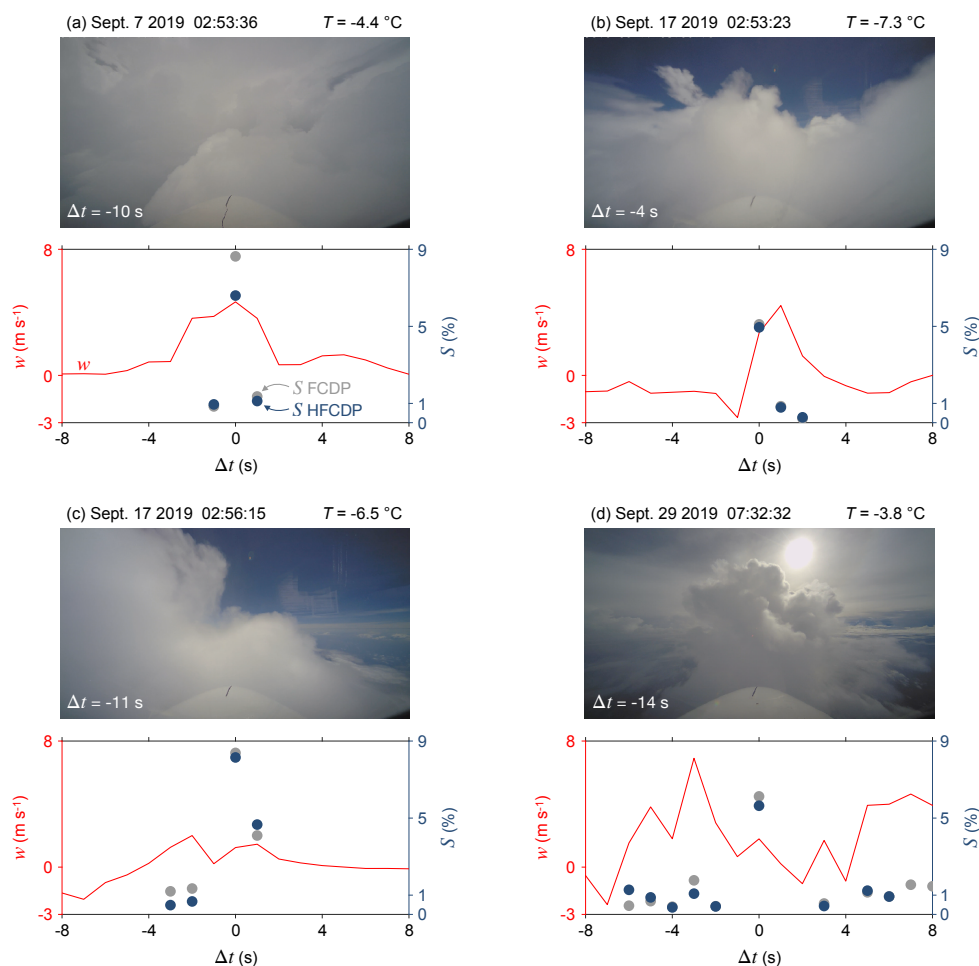


Figure 8: Examination of the seven samples with $S > 5\%$. These samples occurred in four distinct 1-s instances in the Philippines campaign (CAMP²Ex). In each 1-s instance, $S > 5\%$ was estimated with FCDP data, HFCDP data, or both. (a) The first instance occurred on Sept. 7, 2019 at 02:53:36 UTC. The time and temperature of the instance with $S > 5\%$ is printed above the figure. The time lag relative to this instance is denoted by Δt . The top panel shows a representative image from the forward-facing video camera on the aircraft at $\Delta t = -10$ s. The bottom panel shows a timeseries of w and S centered on $\Delta t = 0$ s. S estimates from the FCDP and HFCDP probes are shown separately. (b-d) Similar to (a) for the other three instances in which $S > 5\%$.



3.2 Susceptibility to Warm-Phase Convective Invigoration

We next infer the implications for warm-phase convective invigoration. The analytical model from Part 1 predicts warm-phase invigoration by comparing a clean thermal and a polluted thermal in the same meteorological environment. Invigoration is expressed as a function of the properties of the clean thermal and the difference in cloud-droplet number concentration between the two thermals. Here, we treat observed cloud updrafts as the “clean” thermal, and we use the analytical model to estimate how much faster the updraft would be if the cloud-droplet number concentration were to increase with the meteorological environment held fixed. This approach infers cloud susceptibility to warm-phase invigoration rather than showing evidence of the mechanism itself.

For a given value of raS/c_d , the analytical model predicts the existence of an optimal updraft speed that experiences the largest amount of warm-phase convective invigoration, w^* . The aircraft observations provide a and S . The parameters r and c_d are not readily available from the observations, so they are assumed to be within their typical ranges of 200 to 1000 m and 0.2 to 1, respectively. With these parameter settings, w^* can be estimated for each 1-s cloud sample and compared with the observed updraft speed. We note that by applying the analytical model to each 1-s sample, we are ignoring the difference in horizontal scale between the observations and cloud thermals (~ 100 m and ~ 1000 m, respectively). In other words, invigoration metrics are computed for each ~ 100 m cloud segment assuming that an entire cloud thermal has the properties of that segment. This is expected to overestimate the extreme values of w^* and other invigoration metrics predicted by the model.

One metric of the potential for warm-phase invigoration is the ratio of w^* to w . A ratio of one indicates that the updraft has optimal conditions for invigoration, and a ratio much smaller than one indicates that the updraft would experience a powerful negative feedback from drag that resists invigoration (Part 1). To compare a range of updraft speeds, we composite the observations into w bins with edges at 1, 3, 5, 7, 10, and 15 m s^{-1} . The final bin includes all updrafts that rise faster than 15 m s^{-1} . w^*/w is then computed within each bin assuming that r and c_d are equal to the midpoints of their typical ranges (600 m and 0.6, respectively). Most of the observed updrafts rise faster than the theoretical optimal updraft speed (Fig. 9). The updrafts that are closest to optimal conditions are observed over the ocean with $1 \leq w < 3 \text{ m s}^{-1}$. Under these conditions, the median of $\frac{w^*}{w}$ varies from 0.3 to 0.5 across the different campaigns, indicating that typical updrafts rise at a speed that is within a factor of two or three of optimal conditions. The ratio $\frac{w^*}{w}$ becomes smaller as w becomes faster, and it is smaller over land and coasts than over ocean for a given w bin. The updrafts that are farthest from optimal conditions occur over coasts and land with $w > 15 \text{ m s}^{-1}$. Under these conditions, w is an order of magnitude faster than w^* .

The degree to which updrafts differ from optimal conditions affects the cloud susceptibility to warm-phase invigoration. Let susceptibility be defined by $\Delta \ln w / \Delta \ln n$, where $\Delta \ln n$ is a hypothetical increase in cloud-droplet number concentration and $\Delta \ln w$ is the resulting change in updraft speed caused by warm-phase invigoration. The analytical model predicts susceptibility as a function of raS/c_d and w (Eq. 14 in Part 1). We compute susceptibility assuming $r = 600$ m and

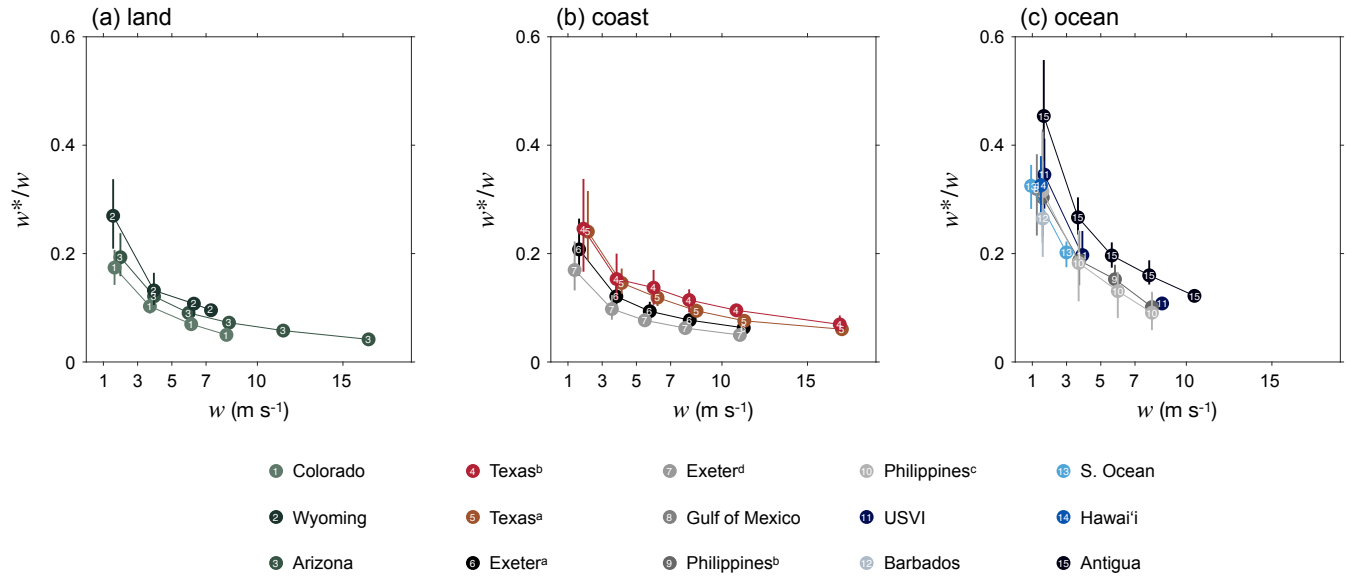


$c_d = 0.6$, as before. The most susceptible updrafts are observed over ocean with $1 \leq w < 3 \text{ m s}^{-1}$ (Fig. 10). Under these conditions, the median of $\Delta \ln w / \Delta \ln n$ varies between 0.11 and 0.25 across the campaigns, implying that updrafts become 1% faster with every 4% to 9% increase in n . Like w^*/w , susceptibility decreases as updrafts become faster, and updrafts over ocean are more susceptible than updrafts over coasts and land for a given w bin. The least susceptible updrafts occur over coasts and land with $w > 15 \text{ m s}^{-1}$. Under these conditions, the median of $\Delta \ln w / \Delta \ln n$ varies between 0.003 and 0.009. This implies that typical updrafts become 1% faster with every 194% to 1815% increase in n .

These susceptibility estimates suggest that warm-phase invigoration may be substantial in some clouds but not others. To quantify this, we consider warm-phase invigoration to be substantial if it causes w to increase by 25% or more in response to a five-fold increase in n , corresponding to $\frac{\Delta \ln w}{\Delta \ln n} \geq 0.14$. The minimum supersaturation required for substantial invigoration, S_{min} , is calculated for each 1-s cloud sample with Eq. 2. Finally, the fraction of updrafts that are susceptible to substantial invigoration, f , is computed as the number of samples with $S \geq S_{min}$ divided by the total number of samples.

We compute f for a range of plausible conditions for the parameters in Eq. 2. For each 1-s sample, one value of r and one value of c_d are randomly selected from uniform distributions spanning 200 to 1000 m and 0.2 to 1, respectively. Data are composited into deciles of w , and f is computed for each w decile. This procedure is repeated 1000 times to estimate uncertainty in f . The central estimate of f is determined by the median of the 1000 realizations, and the 95% confidence interval is determined by the 2.5 and 97.5 percentiles. This confidence interval represents the uncertainty that arises from the fact that r and c_d can be bounded by their typical ranges, but the values for any individual sample are unknown.

f can be decomposed as a function of surface type and updraft speed to identify the conditions that are most and least likely to be susceptible to substantial invigoration (Fig. 11). Oceanic clouds with slow vertical wind speeds have the largest fraction of susceptible updrafts. $74 \pm 3\%$ of oceanic cloud updrafts are susceptible in the lowest w decile, and more than 50% of updrafts are susceptible in the bottom four deciles. The fraction of susceptible updrafts diminishes with each successive w decile, culminating in a value of $f = 1 \pm 1\%$ in the highest decile. Coastal and continental clouds also have the largest fraction of susceptible updrafts in the lowest w decile, but f is smaller than over ocean. Less than 25% of coastal and continental cloud updrafts are susceptible to substantial invigoration in all w deciles, and less than 1% of updrafts are susceptible in the top four coastal deciles and top six continental deciles. Performing the calculations over coarser w bins to synthesize the results, we find that 62% of oceanic updrafts are susceptible to substantial invigoration when $1 \leq w < 2 \text{ m s}^{-1}$, but only 8% are susceptible when $w > 3 \text{ m s}^{-1}$ and only 1% are susceptible when $w > 5 \text{ m s}^{-1}$. Furthermore, 14% of coastal and continental updrafts are susceptible to substantial invigoration when $1 \leq w < 2 \text{ m s}^{-1}$, but only 0.3% are susceptible when $w > 3 \text{ m s}^{-1}$ and only 0.05% are susceptible when $w > 5 \text{ m s}^{-1}$. Similar results are obtained when the observations are horizontally averaged over cloud updrafts prior to computing f (Appendix B). These findings indicate that oceanic cloud updrafts with $w < 2 \text{ m s}^{-1}$ are usually susceptible to substantial warm-phase invigoration, while continental and coastal cloud updrafts with $w > 5 \text{ m s}^{-1}$ are almost never susceptible.



520 **Figure 9: Comparison of observed updraft speed w and theoretical updraft speed for optimal warm-phase invigoration w^* . (a)**
 w^*/w for land campaigns. Data are binned by w according to the values labelled on the horizontal axis. Circles and vertical lines
show the median and interquartile range of w^*/w within each bin. Bins with fewer than eight samples are not shown. (b-c) Similar
to (a), but showing the coast and ocean campaigns, respectively. Superscripts a, b, c, and d indicate that S is estimated using the
CDP, FCDP, HFCDP, and FSSP probes, respectively. w^* is computed assuming that $r = 600$ m and $c_d = 0.6$, where r and c_d are
525 **the radius and effective drag coefficient of the cloud thermal.**

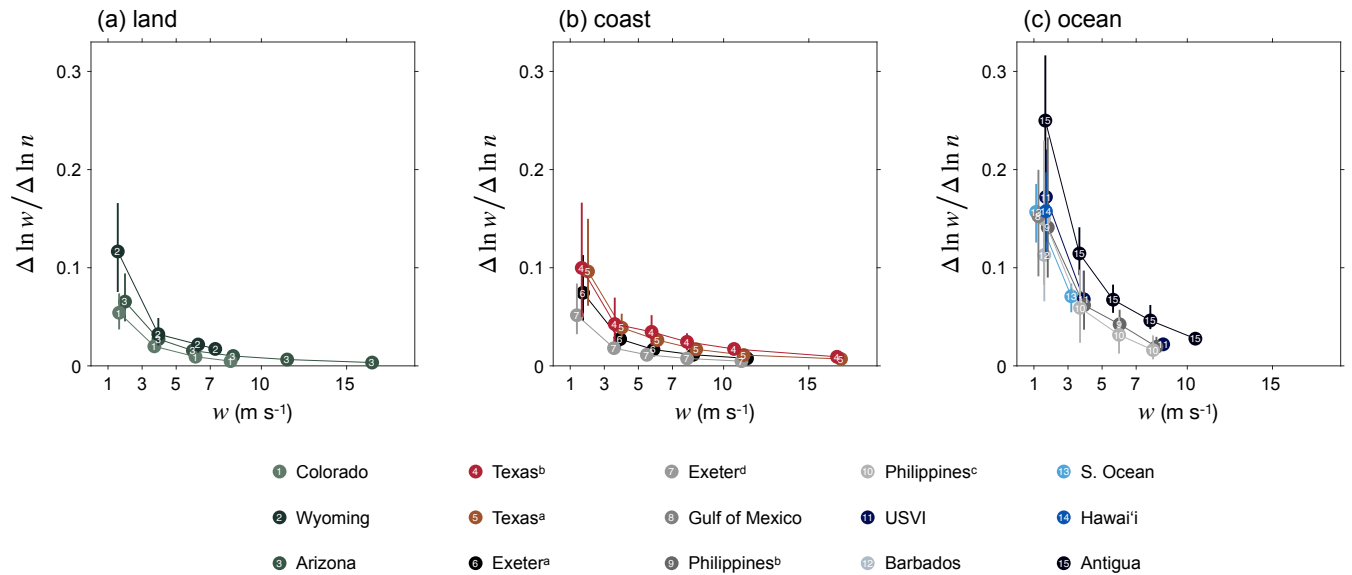




Figure 10: Similar to Fig. 9 but for $\Delta \ln w / \Delta \ln n$.

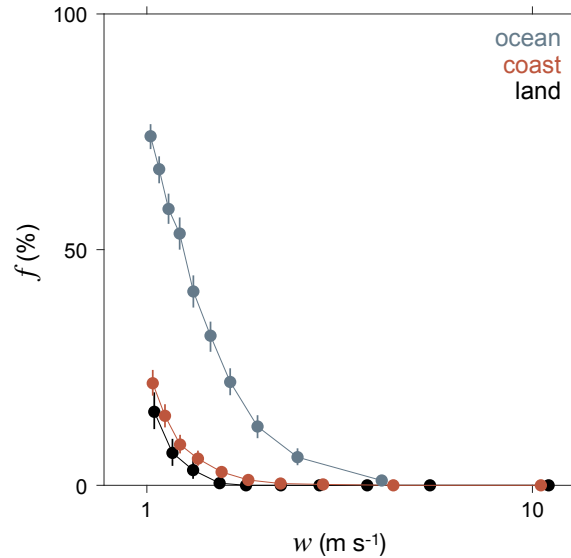


Figure 11: Fraction of cloud updrafts in which S exceeds the theoretical threshold for substantial warm-phase convective invigoration (f). Ocean, coast, and land campaigns are considered separately. In each case, f is plotted for each w decile. Circles and vertical lines show the median and 95% confidence interval of f computed with Monte Carlo sampling of r and c_d (see text).

535 4 Discussion

Our results contribute to a body of literature that uses aircraft observations to estimate S in the liquid layer of convective clouds. Approximately half of these studies reported a handful of cases with $S > 5\%$, and the other half did not (compare Paluch and Knight (1984), Politovich and Cooper (1988), Prabha et al. (2011), Fan et al. (2025), Pan et al. (2026), and Patil et al. (2026) with Warner (1968), Siebert and Shaw (2017), Romps et al. (2023), and Romps (2025a)). Because $S >$
540 5% is a prerequisite for substantial warm-phase invigoration in updrafts speeds typical of oceanic deep convection, a key question is whether these high supersaturations rarely occur, as suggested by some studies, or whether they never occur, as suggested by the others. We address this question by estimating S in 13,342 samples of cloud updrafts at ~ 100 m horizontal resolution. Among these samples, seven have $S > 5\%$ (Fig. 8). Three of these samples may be unreliable because they appear to contain both cloudy and clear air near the cloud top. The four remaining samples include some coincident
545 estimates from two light-scattering probes, so they represent only two separate physical segments of ~ 100 m length. Thus, we find that S almost never exceeds 5% in liquid cloud elements at this length scale.

The existence of $S > 5\%$ in some previous studies can be understood by considering the different filters that have been used to identify valid estimates of S . We demonstrate the implications of filtering by successively applying one



condition after another. Consider a relaxed baseline filter for liquid cloud updrafts defined by (i) the aircraft flies along a flat
550 and straight trajectory when sampling the updraft, (ii) $w > 1 \text{ m s}^{-1}$, (iii) $T > 0 \text{ }^{\circ}\text{C}$ or $\frac{\text{LWC}}{\text{TWC}} > 0.99$, and (iv) $q_t > 10^{-5} \text{ kg kg}^{-1}$
computed with 1-Hz data. This is similar to our filter except that it does not attempt to exclude samples at cloud edges, at
cloud base, or in updrafts that accelerate rapidly. It is also similar to the filters used in some previous studies in the sense that
it identifies liquid cloud updrafts without considering other assumptions on which the quasi-steady approximation is based.
According to this baseline filter, 232 samples have $S > 5\%$, and the 99.9th percentile of S extends to 20% (Fig. 12). If the
555 samples near cloud base are excluded by applying the fifth filtering condition from Sec. 3.1, then 217 of the remaining
samples have $S > 5\%$, and the 99.9th percentile of S increases slightly to 21%. Excluding additional samples at cloud edges
by applying the third condition in Sec. 3.1 leaves 15 samples with $S > 5\%$ and reduces the 99.9th percentile of S to 4.7%.
Finally, excluding samples in rapidly accelerating updrafts by applying the sixth condition in Sec. 3.1 leaves seven samples
with $S > 5\%$ and reduces the 99.9th percentile of S to 3.7%. This final case is equivalent to the filter applied in our analysis.
560 Compared with the baseline, our filter has a sample size that is 43% smaller, but the median and interquartile range of S are
approximately the same. This progression shows that the different filtering choices do not affect statistics for the typical
values of S , but they do affect the extreme upper tail of the distribution. In particular, samples located at cloud edges and in
rapidly accelerating updrafts are unreliable and must be excluded to confidently bound S . Most previous studies did not
systematically exclude all of these samples, with the exception of Romps (2025a), who excluded rapidly accelerating
565 updrafts but not cloud edges, and Patil et al. (2026), who excluded cloud edges but not rapidly accelerating updrafts. The
stringent filtering applied here avoids problematic samples and clarifies the upper bound of S .

Our findings also have implications for understanding which cloud types are most susceptible to warm-phase
convective invigoration. Some of the most prominent evidence in support of warm-phase invigoration has come from
analysis of deep convective clouds (Fan et al., 2018). This has focused previous debates on invigoration of deep convection,
570 while shallow convection has received much less attention (Grabowski and Jarecka (2015) is an exception). Compared to
shallow convection, deep convective updrafts attain larger peak supersaturations because they rise faster and have a deeper
liquid layer that provides more opportunity for droplets to reach the collision-coalescence growth regime. This, in turn,
allows S to change more dramatically in response to a given aerosol perturbation. However, deep convective updrafts also
experience greater resistance to invigoration from stronger drag (Part 1). The competition between these factors determines
575 whether deep or shallow clouds are more susceptible to warm-phase invigoration. Among the clouds sampled in our
analysis, oceanic clouds with $1 \leq w < 3 \text{ m s}^{-1}$ experience a combination of relatively high S and weak drag that makes them
most susceptible to warm-phase invigoration (Fig. 9-11). Thus, in our view, it would be insightful to expand the debate on
warm-phase invigoration by investigating precipitating shallow cumuli over the ocean.

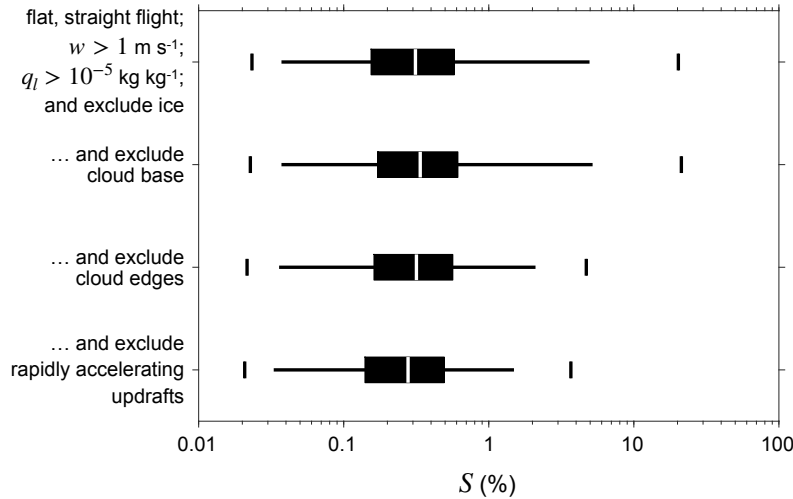


Figure 12: S statistics with different filtering conditions. The topmost case selects updrafts in the liquid part of convective clouds by identifying samples in which the aircraft flies along a flat and straight path, $w > 1 \text{ m s}^{-1}$, the cloud liquid mixing ratio q_l is greater than $10^{-5} \text{ kg kg}^{-1}$, and cloud ice is negligible. The remaining cases add other filtering conditions one at a time, including the exclusion of samples near the cloud base, exclusion of samples near cloud edges, and exclusion of samples in rapidly accelerating updrafts (see text). Conditions are added successively moving downward on the plot, so each filter is more restrictive than the one above it. The lowermost filter is used in the main analysis. Box plots show the 1, 25, 50, 75, and 99 percentiles, and the outer vertical lines show the 0.1 and 99.9 percentiles. The sample sizes for the different filters, ranked from least to most restrictive, are 23,516; 20,643; 17,830 and 13,297. Statistics are computed by compiling data from all campaigns except for the one over the Gulf of Mexico (SEAC⁴RS). This campaign is excluded because it does not provide the necessary data to compute the final filtering condition.

5 Conclusion

This two-part study investigates warm-phase convective invigoration from aerosol-cloud interactions with theory and observations. Part 1 derives an analytical model that predicts how much warm-phase invigoration changes convective updraft speeds if the cloud-droplet number concentration increases and the meteorological environment is fixed. Invigoration is considered to be substantial if it causes w to increase by 25% or more when the cloud-droplet number concentration increases by a factor of five. This analytical model makes two key predictions. First, for cloud thermals with a given S , T , p , r , and c_d , there exists an optimal updraft speed that experiences the largest amount of warm-phase invigoration, w^* . Second, substantial invigoration can occur only if S exceeds a threshold S_{min} that depends on the properties of the thermal. These predictions raise two questions: Under what circumstances are convective updraft speeds close to w^* , and how often do convective updrafts have $S > S_{min}$?

Part 2 of this study investigates these questions using aircraft observations that measure the liquid layer of convective clouds. The primary challenge of the observational analysis is to quantify S with a method that is accurate and



includes a wide range of convective environments. We address this challenge by analyzing 150 research flights from 12
605 different aircraft campaigns that are distributed across the globe. The flights measure deep and shallow convection in a
variety of environments over land, coasts, and ocean. The quasi-steady approximation is applied to estimate S in liquid
elements of convective updrafts with a horizontal resolution of ~ 100 m. Stringent filtering criteria are developed to exclude
samples in which the assumptions of the quasi-steady approximation are not satisfied, including samples at cloud base, at
cloud edges, and in rapidly accelerating updrafts. The resulting database includes 13,342 quality-controlled samples of S .
610 Collectively, these data have sufficient accuracy, reproducibility, and geographic distribution to draw strong conclusions
about warm-phase invigoration (Fig. 1, 2).

The first question is investigated by comparing the observed updraft speed to w^* . Most of the observed updrafts rise
faster than w^* (Fig. 9). The updrafts that are closest to optimal conditions are observed over the ocean and have $1 \leq w < 3$
 m s^{-1} . Under these conditions, w is typically between two and three times faster than w^* . The ratio w^*/w progressively
615 decreases as w becomes faster, and it is smaller over coasts and land than over ocean for a given w . The clouds that are
farthest from optimal conditions are observed over coast and land have $w > 15 \text{ m s}^{-1}$. Under these conditions, w is an order
of magnitude faster than w^* .

The second question is investigated by comparing the observed values of S to the theoretical values of S_{min} . The
fraction of samples with $S > S_{min}$ is computed to quantify the fraction of cloud updrafts that are susceptible to substantial
620 warm-phase invigoration. 62% of oceanic cloud updrafts are susceptible to substantial invigoration when $1 \leq w < 2 \text{ m s}^{-1}$,
but only 8% are susceptible when $w > 3 \text{ m s}^{-1}$ and only 1% are susceptible when $w > 5 \text{ m s}^{-1}$ (Fig. 11). 14% of coastal and
continental cloud updrafts are susceptible to substantial invigoration when $1 \leq w < 2 \text{ m s}^{-1}$, but only 0.3% are susceptible
when $w > 3 \text{ m s}^{-1}$ and only 0.05% are susceptible when $w > 5 \text{ m s}^{-1}$. These results indicate that precipitating shallow cumuli
625 over the ocean are usually susceptible to substantial warm-phase invigoration, while deep convective clouds over coast and
land are almost never susceptible. Given that recent debates on warm-phase invigoration have focused on deep convection,
an important next step is to better understand the mechanism in oceanic shallow cumuli.

Appendix A: Additional Information about the Observations

Additional information about the radar resolution, radiosonde locations, and excluded research flights are presented
in Tables A1, A2, and A3, respectively.

630

Campaign	Temporal Resolution (s)	Vertical Resolution (m)	Blind Range (m)
CuPIDO	0.04	typically 45 but varies from 15 to 75	105
SPICULE	0.1	19	107



CloudGPS	0.03	30	75
COPE	0.06	15	105
ESCAPE	0.08	typically 34 but varies from 17 to 34	103
SEAC ⁴ RS	-	-	-
ICE-T	0.1	15	105
RICO	0.12	typically 45 but varies from 15 to 75	105
EUREC ⁴ A	2	30	90
CSET	0.1	19	107
CAMP ² Ex	1.8	30	390
SOCRATES	0.5	19	107

Table A1: Nominal resolution and blind range of the radar data products. The blind range is the vertical distance between the aircraft and the closest range bin with available data.

Campaign	Radiosonde Station	Station ID	References
CuPIDO	Tucson, Arizona, USA	72274 TUS	University of Wyoming (2006a), University of Wyoming (2006b)
SPICULE	Grand Junction, CO, USA	72476 GJT	University of Wyoming (2021a), University of Wyoming (2021b)
CloudGPS	North Platte, NE, USA	72562 LBF	University of Wyoming (2008)
COPE	Camborne, UK	03808	University of Wyoming (2013a), University of Wyoming (2013b)
ESCAPE	Lake Charles, LA, USA	72240 LCH	University of Wyoming (2022)
SEAC ⁴ RS	Lake Charles, LA, USA	72240 LCH	University of Wyoming (2013c), University of Wyoming (2013d)
ICE-T	San Juan, Puerto Rico	78526 TJSJ	University of Wyoming (2011)
RICO	San Juan, Puerto Rico	78526 TJSJ	University of Wyoming (2004), University of Wyoming (2005)
EUREC ⁴ A	Grantley Adams airport, Barbados	78954 TBPB	University of Wyoming (2020a), University of Wyoming (2020b)



CSET	Hilo, HI, USA	91285 PHTO	University of Wyoming (2015a), University of Wyoming (2015b)
CAMP ² Ex	Laoag, Philippines	98223 RPLI	University of Wyoming (2019a), University of Wyoming (2019b), University of Wyoming (2019c)
SOCRATES	Invercargill, NZ	93844 NZNV	University of Wyoming (2018a), University of Wyoming (2018b)

Table A2: Locations of soundings from the University of Wyoming Atmospheric Science Sounding Archive.

635

Campaign	Excluded Research Flights	Explanation
EUREC ⁴ A	RF02-RF10, RF20	Radar not available or bad in-situ vertical wind speed data (Bony et al., 2022)
CAMP ² Ex	August 30	Vertical wind speed data not available
CloudGPS	All except May 21, 23, and 26	Plane was struck by lightning during May 26 flight, and FSSP was not reliable afterwards according to flight notes. Radar not available on May 19 flight.
CuPIDO	July 18	Radar not available
ESCAPE	RF01 and RF07	Radar not available
ICE-T	RF02-RF07	Radar not available
RICO	December 9	Imaging probe not available
SEAC ⁴ RS	All except August 12, September 4, September 13, and September 18	Included flights sampled convective clouds over the Gulf of Mexico. Excluded flights sampled other conditions over continental USA.

Table A3: Summary of research flights that were excluded from the analysis.

Appendix B: Sensitivity Tests

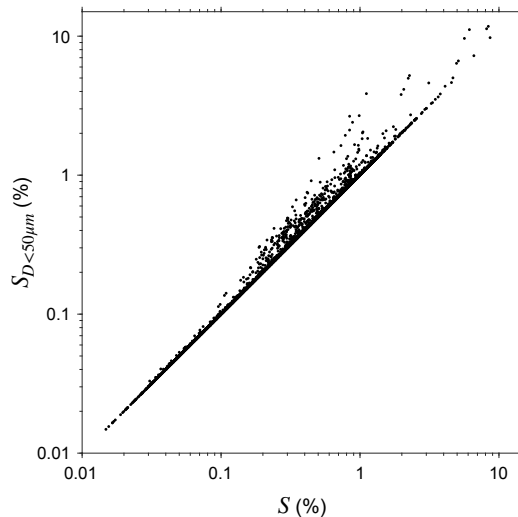
Here we present two sensitivity tests. The first test shows how large hydrometeors affect the estimated supersaturation, and the second test shows how horizontal averaging affects the conclusions.

The first sensitivity test checks whether S can be estimated accurately when an imaging probe is not available. The S estimates depend on the sum of the cloud-droplet diameters per unit volume of air, expressed as $n\bar{D}$ (Eq. 1). In most cases, $n\bar{D}$ is computed using an imaging probe to measure the particles with $D \gtrsim 50 \mu\text{m}$. We check how much the S estimates change if these measurements are excluded, thereby truncating the hydrometeor size distribution at an upper bound of $D \approx 50 \mu\text{m}$. This approximation is made in the Texas campaign (ESCAPE) because a quality-controlled scientific dataset from



the imaging probe was not available at the time of study. Let S and $S_{D<50\mu m}$ be the estimated supersaturation including and excluding the imaging probe, respectively. Across all campaigns except for the one over Texas (ESCAPE), a fraction of 0.92 of the samples have $|S - S_{D<50\mu m}| < 0.01\%$, a fraction of 0.05 of the samples have $0.01\% \leq |S - S_{D<50\mu m}| < 0.1\%$, and the remaining fraction of 0.02 of the samples have $|S - S_{D<50\mu m}| \geq 0.1\%$ (Fig. B1).

650 The second sensitivity test checks how horizontal averaging affects the fraction of cloud samples with $S > S_{min}$. Fig. 11 presents f computed with observations that have ~ 100 m horizontal resolution. Here, we perform a similar calculation at the horizontal scale of cloud updrafts. Let cloud updrafts be defined as contiguous groups of points in which $w \geq 1 \text{ m s}^{-1}$ and $q_l \geq 10^{-5} \text{ kg kg}^{-1}$, and let overbars denote horizontal averages over cloud updrafts. The fraction of observations with $\bar{S} \geq S_{min}$ in a given $\bar{w}$ decile takes on similar values to the corresponding w decile in Fig. 11 (Fig. B2).
655 This means that the main conclusions about susceptibility to substantial invigoration are similar whether horizontal averaging is performed over ~ 100 m segments or over entire cloud updrafts.



660 **Figure B1: Sensitivity test showing how much the supersaturation estimates change if hydrometeors with diameters greater than $\sim 50 \mu m$ are excluded from the calculation. The supersaturation computed while ignoring large hydrometeors, $S_{D<50\mu m}$, is plotted as a function of the supersaturation computed while considering all hydrometeors, S . Data from all campaigns except for the one over Texas (ESCAPE) are included. This campaign is excluded because a quality-controlled scientific dataset for hydrometeors with diameters larger than $50 \mu m$ from an imaging probe was not available at the time of study.**

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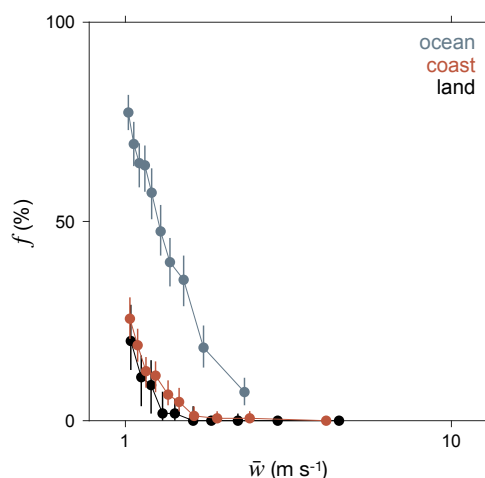


Figure B2: Similar to Fig. 11 except that aircraft observations are horizontally averaged over cloud updrafts prior to computing f . Overbars indicate horizontal averages over cloud updrafts.

Code Availability

670 Matlab code to reproduce the figures in the manuscript is available from the corresponding author upon request.

Data Availability

CuPIDO data are available from https://data.eol.ucar.edu/master_lists/generated/cupido/. SPICULE data are available from https://data.eol.ucar.edu/master_lists/generated/spicule/. CloudGPS data are available from the data curators upon request from <https://www.uwyo.edu/atsc/research-facilities/uwka/instruments/data-request.html>. ESCAPE data are available from https://data.eol.ucar.edu/master_lists/generated/escape/. SEAC⁴RS data are available from <https://www-air.larc.nasa.gov/cgi-bin/ArcView/seac4rs>. COPE data are available from https://www.uwyo.edu/atsc/_files/uwka/projectarchive/2013-cope-med.pdf. ICE-T data are available from https://data.eol.ucar.edu/master_lists/generated/ice-t/. RICO data are available from https://data.eol.ucar.edu/master_lists/generated/rico/. EUREC⁴A data are available from <https://safireplus.aeris-data.fr/data-access/?tab=1&id=EUREC4A>. CSET data are available from https://data.eol.ucar.edu/master_lists/generated/cset/.
680 CAMP²Ex data are available from <https://www-air.larc.nasa.gov/cgi-bin/ArcView/camp2ex>. SOCRATES data are available from https://data.eol.ucar.edu/master_lists/generated/socrates/. Data from the University of Wyoming radiosonde archive are available from https://weather.uwyo.edu/upperair/sounding_legacy.html. MERRA-2 data are available from the NASA Goddard Earth Sciences Data and Information Services Center (<https://doi.org/10.5067/LTVB4GPCOTK2>, NASA, 2023).



Author Contributions

685 Conceptualization: CJW, ACV, RÖ, DMR; Formal Analysis: CJW; Funding Acquisition: CJW; Methodology: CJW, DMR; Project Administration: CJW; Software: CJW; Visualization: CJW; Validation: RÖ; Writing, Original Draft: CJW; Writing, Review and Editing: CJW, ACV, RÖ, DMR.

Competing Interests

The authors declares that they have no conflict of interest.

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