# Arabian Sea tropical cyclones intensified by emissions of black carbon and other aerosols

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Conditionally accepted: *Nature***Throughout the year, average sea surface temperatures in the Arabian Sea are warm enough to support the development of tropical cyclones1, but the atmospheric monsoon circulation and associated strong vertical wind shear limits cyclone development and intensification, only permitting a pre- and post-monsoon period for cyclogenesis1-4. Thus a recent increase in the intensity of Northern Indian Ocean tropical cyclones5 is thought to be related to the weakening of the climatological vertical wind shear3,4.At the same time, anthropogenic emissions of aerosols have increased 6-fold since the 1930s, leading to a weakening of the southwesterly lower-level and easterly upper-level winds that define the monsoonal circulation over the Arabian Sea6-9.In principle, this aerosol-driven circulation modification could affect Arabian Sea tropical cyclone intensity, but to date, no such linkage has been shown.Here we report an increase in the intensity of pre-monsoon Arabian Sea tropical cyclones over the past 30 years, and demonstrate thatthis change in storm strength is a consequence of a simultaneous upward trend in anthropogenic black carbon and sulfate emissions.We use a combination of observational, reanalysis, and model data to demonstrate that the anomalous circulation, which is radiatively forced by these anthropogenic aerosols, reduces the basin-wide vertical wind shear, creating an environment more favourable for tropical cyclone intensification.As most Arabian Sea tropical cyclones make landfall1 our results suggest an additional human health impact from regional air pollution.**

The South Asian atmospheric brown cloud (ABC) is a 3-km thick layer of pollution over the Northern Indian Ocean and Indian subcontinent that results from human emission of aerosols like black and organic carbon carbon and sulfates9. When over water the dominant surface radiative effect of the ABC is a reduction in solar insolation, which is enhanced significantly by black and organic carbon aerosols7-11. There is a northward gradient of fine mode aerosol optical depth over the Arabian Sea (Fig 1a)and therefore a southward gradient in aerosol surface forcing7,9,10,12 (i.e., negative surface forcing is greater in magnitude to the north). Positive forcing associated with greenhouse gasses is offset in the NIO by the 6-fold increase in emissions since 19309, resulting in a southward gradient in sea surface temperature (SST) trends (Fig 1b) and a reduction in the climatological summertime northward SST gradient7. The atmospheric response to this anomalous SST gradient is a weakening of the southerly cross-equatorial surface flow and upper level tropical easterly jet, which define the Indian monsoon circulation, via anomalous high (low) sea level pressure to the north (south)7,8.

Once the monsoon onset occurs very strong vertical wind shear develops across the Arabian Sea, the main factor prohibiting tropical cyclone development during the months of July and August1,2. Thus, despite climatologically warm SST, only two to three tropical cyclones form in the Arabian Sea every calendar year, an average of one storm in the pre-monsoon period (May-June), and one to two storms in the post-monsoon period (August–December)1.

For a tropical cyclone to intensify net heat transfer from the ocean to the near surface air must be sufficient to overcome the frictional drag acting on the cyclone winds in the boundary layer13. In the absence of large-scale kinematic forces that would disrupt the physical structure of a cyclone, such as vertical wind shear14, storm intensity is limited largely by the thermodynamic environment, as described by potential intensity theory15. As atmospheric kinematic conditions become more favourable for cyclones to develop those that do form are more likely to achieve maximum intensities close to their potential intensity. Note that the ABC can affect the mean thermodynamic environment by stabilizing regional atmospheric lapse rates and reducing potential intensity, but this effect is small relative to the regionally high absolute values of potential intensity1.

One way to evaluate temporal changes in the intensity of tropical cyclones is to examine cyclone lifetime maximum intensity (LMI), defined as the maximum intensity achieved in the lifetime of a storm16. We calculate LMI for each storm from historical cyclone wind speed estimates17,18 and then separate the data into the 1979–1996 and 1997–2010 periods, as 10 pre-monsoon tropical cyclones formed during each epoch. The distribution of pre-monsoon LMI shifts substantially toward greater intensity from the early to later period while the post-monsoon distribution shifts slightly downward (Fig 2a). The difference in the pre-monsoon median (and mean) LMI between the periods is significant at greater than the 97% level, but there is no significant statistical separation for the post-monsoon storms.

Arabian Sea tropical cyclone intensity estimates are based on Dvorak analysis of satellite imagery19, and given improvements in observational platforms over time it is plausible that the LMI of pre-1998 storms is underestimated, thus causing an apparent increase in the LMI of storms forming after this time. However, if this were the case we would expect to see an increase in the LMI of post-monsoon storms since there is no physical reason why such data artifacts would only affect the pre-monsoon cyclones. Therefore, the striking dissimilarity between the pre- and post-monsoon distribution changes suggests the change in the pre-monsoon LMI distribution is physical. We repeated this analysis with intensity estimates from an objective analysis of satellite data that is corrected for artifacts related to temporal changes in the observational platform16,20, obtaining nearly identical results (not shown).

We interpret the increasing LMI of pre-monsoon cyclones as reflecting a more favourable kinematic environment for storms to intensify, especially as persistently high SST helps to maintain a thermodynamic environment that is favourable for storm intensification1,3,4. We hypothesize that a reduction in the environmental vertical wind shear (defined here as the magnitude of the vector difference between the upper- and lower-level wind measured along the 200 and 850 mb pressure levels, respectively) that the storms are experiencing is the catalyst for the increase in pre-monsoon LMI1,3,4.

Using reanalysis data21 we test this hypothesis by calculating the storm-ambient vertical wind shear, defined as the vertical wind shear at the location of the storm averaged over the period during which the tropical cyclone’s intensity increases from 17 m s-1 to its LMI. Similar to the procedure of Kossin and Camargo22, we averaged vertical shear values for every cyclone fix 48 hours prior to the arrival of the storm to reduce contamination of the reanalysis fields by the storms themselves. The distribution of pre-monsoon storm-ambient shear exhibits a pronounced shift towards lower values from the earlier to the latter period (Fig 2b); the median shear value decreases from 11 to 8 m s-1, and the lower quartile for the early period is equal to the upper quartile of the latter period.

Models23 and observations24 suggest that the relationship between vertical wind shear and tropical cyclone intensity is not linear, rather a tropical cyclone will tend to weaken (strengthen) in the presence of vertical shear greater than (less than) values within the approximate range of 8-11 m s-1. Therefore, as the pre-monsoon shear values become more likely to fall below this range the storms are more likely to continue intensifying toward their potential intensityand their LMI distribution is expected to shift toward higher values. Although the post-monsoon storm-ambient shear distribution also shifts toward lower values, these values remain mostly within or above this range of thresholds for intensification, and the 1997–2010 post-monsoon shear distribution is statistically not different from the 1979–1996 pre-monsoon shear distribution (Table S1). We repeated this analysis using two additional data sets (Figs S3–S5), obtaining quantitatively similar results (Table S1).

We propose that this change in storm-ambient shear is related to a broader downward trend in regional vertical wind shear that results from a simultaneous increase in regional anthropogenic emissions of aerosols. Observational evidence demonstrates that the ABC leads to widespread surface solar dimming of the order of 15 to 20Wm-2 over the Arabian sea during the pre-monsoon months10, and trend analyses of surface solar radiation observations by broad band solar radiometers9,25 reveal a nearly 20 Wm-2 reduction in surface solar radiation from 1950 to 2000, approximately 10% of the climatological solar radiation absorbed at the surface.The dimming is accompanied by a large solar heating of the atmosphere, thus stabilizing the column and further exacerbating the weakening of the circulation resulting from the dimming9. We quantify the influence of ABC surface and atmospheric radiative forcing on vertical wind shear via numerical experiments designed to identify the response of the regional circulation to the increasing emissions of aerosols over the period of 1950-2000(model experiment details are in Methods).

We estimate the 30-year change in observed pre-monsoon SST by multiplying by 30 the averaged linear trend of observed May and June monthly mean SST26 over the period 1979–2010. To facilitate comparison of SST between the model output and observations we subtract from the observed 30-year SST change the equatorial mean (5°S-5°N & 55°-75°E) 30-year change in SST of 0.3°C (Fig 3a). We interpret this equatorial change in observed SST as reflecting positive forcing from greenhouse gasses. The anomalous 30-year SST change from the model experiments is obtained by differencing the May-June ensemble mean of the perturbation and control experiments. Since the model experiments reflect the SST change from the growth of the optical depth of the ABC over the 50-year period of 1950-2000 we assume a linearized response of SST to ABC radiative forcing and scale the model results by ⅗ to estimate the SST change from a 30-year increase in the optical depth of the ABC (Fig 3b). The observed and modeled southward gradient of the 30-year SST change are similar in magnitude, with SST warming at 5°N being 0.5°C greater than the warming at 15°N in both.

We estimate the historical 30-year change in pre-monsoon vertical wind shear by multiplying by 30 the averaged May and June linear trend of vertical shear from reanalysis21 (Fig. 3a). We assume a linear response of the atmosphere to growth of the ABC and estimate the ABC-forced 30-year change in pre-monsoon vertical wind shear as the difference in ensemble mean May-June vertical shear, scaled by ⅗ (Fig. 3b). The reanalysis and modeled 30-year change in pre-monsoon vertical wind shear are in good agreement north of 10°N, where the ABC is present and where the majority of pre-monsoon Arabian Sea tropical cyclones form and track (Fig. 1a). Each data exhibit a 1.5 m s-1 reduction off the western coast of the Indian Subcontinent, and northward and westward from this location the change in shear in each is 1.0 m s-1. This agreement is compelling evidence that the basin-wide reduction in vertical wind shear is causally related to the increasing intensity of the ABC. We compared the model output with trends in vertical shear from two additional reanalysis data sets, obtaining nearly identical results (Fig S6).

Noting that the distribution of post-monsoon storm-ambient shear and LMI during the 1997–2010 period is similar to those pre-monsoon distributions for the 1979–1996 time span (Fig 2, Table S1),it can be arguedthat further reduction of basin-wide shear associated with projected growth in emissions9will further shift the inner quartile range of the post-monsoon storm-ambient shear distribution toward that of the pre-monsoon shear during the latter period. In such a case it is plausible that very intense tropical cyclones, which have thus far been limited to the pre-monsoon period, could begin to emerge in the post-monsoon season as well.

Given the relatively small size of the Arabian basin over half of all cyclones that form here make landfall, and even weak landfalling Arabian Sea cyclones can cause considerable destruction and loss of life1,18,27. All of the most powerful Arabian Sea cyclones that occurred during the last 30 years formed during the pre-monsoon period and made landfall in either India, Pakistan, Oman, or Iran, causing considerable loss of life and substantial damages (Table 1). In addition to the multitude of known human health impacts associated with aerosols that comprise the ABC28 we suggest that the increasing intensity of landfalling tropical cyclones is a consequence of regional emissions of pollution aerosols. Since tropospheric aerosols have a short residence time, reducing emissions should have a nearly immediate effect on the propensity of pre-monsoon tropical cyclones to reach their maximum potential intensity. However, we caution that this study is conducted from a limited sample of storms and continued research that tests the relationship between aerosols and tropical cyclones in the Northern Indian Ocean is warranted.

# Methods Summary

The numerical experiments used here are identical to those described in *Chung and Ramanathan*7, and we refer the reader to this paper and the online Methods for a detailed discussion of the model setup. These experiments were performed using an atmospheric general circulation model29. In the control and perturbation experiments an SST annual cycle is prescribed. In the control experiment SST is prescribed as observed mean SST26 changes from 1950-2000. In the perturbation experiment SST in the NIO is estimated to be the equivalent 50-year change in SST assuming regional emissions had been stabilized at 1950s levels,so that SST is everywhere the same as in the control experiment, except in the NIO, where the local 50-year SST change is forced to be equal to the SST change in the equatorial Indian Ocean9.SST in the NIO is warmer in the perturbation run than it is in the control run, and the difference in the zonally-averaged (Indian Ocean) meridional SST gradient in the experiments is nearly identical to that from Figure 1b, except that in the model the difference is zero at the equator and negative to the north. The perturbation experiment also includes the seasonally varying effect of heating in the atmosphere by the ABC30. We interpret the difference between the 50-member ensemble mean of the control and perturbation experiments as the influence of the 50-year growth in regional emissions on the atmospheric circulation. References

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# Acknowledgements

Partial funding for this work was provided by NOAA/CPO (NA10OAR4310136), Korea's Research Agency for Climate Science (RACS 2010-2603), and the National Science Foundation (ATM-0721142). NCEP Reanalysis 1 and 2 data provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their web site at http://www.esrl.noaa.gov/psd/.

# Author Contributions

A. E. conceptualized the project, A. E. and J. K. designed the study, A. E., J. K., and E. C. provided model and observational data. All authors participated in data interpretation and co-wrote the manuscript.

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Tables

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| --- | --- | --- | --- | --- | --- | --- |
| Cyclone Name | Year | LMI (m s-1) | Landfall Location | Loss of life | Total Affected | Damages  (Million 2011 USD) |
| 03A | 1998 | 54 | Gujarat India | 2871 | 2871 | 643 |
| 02A | 1999 | 57 | Karachi Pakistan | 682 | 576,636 | 42 |
| 01A | 2001 | 57 | Gujarat India | None | None | None |
| Gonu | 2007 | 75 | Muscat Oman & Ras Al Kuh Iran | 88 | 180,009 | 4203 |
| Phet | 2010 | 64 | Muscat Oman &Karachi Pakistan | 39 | 4000 | 1845 |

**Table 1 Name, year, LMI, and damagesfrom the five most powerful Arabian Sea tropical cyclones.**Shown are data for the five most powerful Arabian Sea cyclones over the last 30 years, all of which formed during the pre-monsoon period. Landfall locations, fatalities, numbers affected, and damages data are from the Joint Typhoon Warning Center‘s Annual Tropical Cyclone Reports18 and the International Disaster Database27.

Figure Legends

**Figure 1. Arabian Sea tropical cyclone tracks, aerosol optical depth, and meridional SST trends.** Genesis points (circles) and tracks (solid lines) are of pre-monsoon tropical cyclones during the period 1979–2010 (a). Storms with an LMI >50 m s-1 are indicated with a filled circle at the genesis point and thick track lines. Shaded contours represent annual long term mean fine mode aerosol optical depth (AOD) from the MODIS Terra and Aqua instruments averaged over 2003–2009. The 50-year change in observed SST, averaged over 55°–75°E (b). The SST change is defined as the average of the monthly linear trend, from 1955–2004, multiplied by 50.

**Figure 2.Distributions of pre- and post-monsoon LMI and storm-ambient vertical wind shear.**In the box plotsof LMI (a) and storm-ambient vertical wind shear (b) shown the median (red line), inner quartile range (blue box), and the 25th (75th) percentile minus (plus) 1.5 times the inner quartile range (horizontal black line). Shear is calculated from the National Center for Environmental Prediction-Department of Energy Reanalysis21. Significance of the separation of the median and mean values is in Table S1.

**Figure 3. 30-year trends in pre-monsoon SST and vertical wind shear.** Plotted is the 30-year trend in vertical wind shear (contours) based on reanalysis data from 1979–2010 (a) and numerical experiments designed to isolate the effect of the ABC on the regional circulation7 (b). Dashed contours indicate negative trends and solid contours are positive, the zero contour is not shown. Positive and negative shear contours are in units of 0.5 m s-1. Shaded are the 30-year SST trendsover the same period from observations (a), which is relative to the equatorial SST trend, and the aerosol-forced SST change prescribed in the model experiments (b).

# Methods (online)

The numerical experiments used in this study to identify the effect of increasing emissions on vertical wind shear are identical to those described in *Chung and Ramanathan*7, and we refer the reader to this paper for a comprehensive discussion of the experimental setup and model results. These numerical experiments were performed with the National Center for Atmospheric Research Community Climate Model Version 3 (NCAR/CCM3)29 at a T42/L18 resolution with prescribed SST. For each of the control and perturbation experiments there were 10 ensemble member simulations, all 10 with different initial conditions. For each of the 10 ensemble simulations the model was spun up to equilibrium for one year, and then run for an additional five years. The ten 5-year simulations were averaged yielding a 50-year sample for each of the ensemble means considered here.

In the control run and perturbation run an SST annual cycle is prescribed. In the control experiment SST is prescribed to globally reflect observed long-term SST26 changes from 1950-2000. This is achieved by calculating the monthly linear SST trend over the period of 1951-2002, multiplying the trend by 50 (°C 50-yr-1), and adding this 50-year SST change to the observed monthly SST climatology (calculated over the 1951–2002 period). Therefore, in the control experiment, SST in the NIO reflects a 50-year increase in emissions and negative surface forcing by the ABC. In the perturbation experiment SST in the NIO is estimated to be the equivalent 50-year change in SST assuming regional emissions had been stabilized at 1950s levels, such that SST is everywhere the same as in the control experiment, except in the NIO, where the local 50-year SST change is forced to be equal to the SST change in the equatorial Indian Ocean, consistent with the findings of *Ramanathan et al.*9.

SST in the NIO is warmer in the perturbation run than it is in the control run, and the difference in the zonally-averaged (Indian Ocean) meridional SST gradient in the experiments is nearly identical to that from Figure 1b, except that in the model the difference is zero at the equator and negative to the north. This difference in SST is equivalent to the radiatively forced change in SST from the increase in emissions over the same 50-year period, as demonstrated by *Ramanathan et al.*10. The perturbation experiment also includes the effect of heating in the atmosphere by the absorption of short and longwave radiation by the ABC; consistent with *Chung and Ramanathan*30, a reduction in surface solar insolation was prescribed, as was atmospheric heating from the surface boundary layer to roughly 3 km height. The seasonality of the surface forcing and imposed heating is consistent with that of the ABC9. We interpret the difference between the ensemble mean of the control and perturbation experiments as the influence of the 50-year growth in regional emissions on the atmospheric circulation.