Cassini Imaging Science Subsystem observations of Titan’s south polar cloud

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\textbf{A B S T R A C T}

In May of 2012 images of Titan obtained by the Cassini Imaging Science Subsystem (ISS) showed a newly-formed cloud patch near the southern pole. The cloud has unusual morphology and texture suggesting that it is formed by condensation at an altitude much higher than expected for any of the known organics in Titan’s atmosphere. We measured the altitude to be 300 ± 10 km from images when the feature was on the limb. Limb images suggest that the initial stages of the formation began in late 2011. It was just visible in images obtained in 2014 but is not expected to be visible in the future due to enveloping darkness as the season progresses. The feature has a slightly different color than the surrounding haze. Its optical thickness is near 2 at 889 nm wavelength and the particle imaginary refractive index must be less than $5\times10^{-4}$ at that wavelength. Wind vectors derived from a time series show that it is rotating about a center offset by 4.5° from Titan’s solid-body spin axis, consistent with that found from the temperature field by Achterberg et al. (Achterberg, R.K., Conrath, B.J., Gierasch, P.J., Flasar, F.M., Nixon, C.A. [2008a]. Icarus 197, 549–555) and subsequent measurements. The feature rotates at an angular velocity near the rate expected for transport of angular momentum from the low latitudes to the pole. The clumpy texture of the feature resembles that of terrestrial cloud fields undergoing open cell convection, an unusual configuration initiated by downwelling.

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1. Introduction

In May of 2012, after 82 close encounters with Saturn’s moon Titan over a period of nearly 8 years, a new and unexpected feature appeared in Cassini Imaging Science Subsystem (ISS) images in the form of a cloud patch at high southern latitudes. By its morphology, texture, and apparent altitude it was clear that nothing similar had been seen on Titan before. For reasons discussed below it appeared to be a condensate forming at an altitude much higher (and therefore much warmer) than expected for any condensable species known to be present. At the same time observations by the Cassini Infrared Spectrometer (CIRS) instrument revealed changes in the distribution of temperature and chemical species driven by seasonal change. Nearly coincident with the discovery of the cloud patch in ISS images, the CIRS instrument also recorded for the first time at high southern latitudes a spectral feature at 220 cm$^{-1}$ that had previously been seen only in the northern winter high latitudes and was thought to be the signature of a condensate of unknown composition (Jennings et al., 2012).

We present observations of the cloud patch made by the ISS instrument from its inception to early 2014. In Section 2 we discuss polar haze enhancement as seen in the limb view where sensitivity to small change is greatest. Limb profiling also provides an accurate measure of the altitude. In this section we also present the earliest observations of enhanced haze opacity at high southern latitudes in an attempt to trace its origins, important for studies of the early formation process and relation to changes occurring in the chemistry and temperature fields. In Section 3 we remark on color and on color difference with respect to the nearby polar haze. We computed radiative transfer models to derive cloud optical depth and absorption and those results appear in Section 4. We also measured wind vectors, presented in Section 5. In Section 6 we discuss the implications of our observations for atmospheric dynamics and relation to chemical and temperature fields.

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2. Limb and near-limb images

Unambiguous evidence for enhancement of Titan’s south polar haze came first from limb images against Saturn’s illuminated disk taken on 6 May (DOY 127) of 2012, hereafter referred to in the format 2012-127 (Fig. 1).

Close inspection of the images from 2012-127 (6 May) reveals a ‘detached haze’ layer that extends over all latitudes south of about 55°N and an enhanced region near the south pole. The south polar features are difficult to discern in the near-ultraviolet because the line-of-sight optical depth of the haze near 300 km altitude is significant, whereas in the near-infrared it is much smaller. This strong dependence on wavelength is characteristic of Titan’s haze in the region around 300 km and somewhat deeper and is interpreted as a signature of the small primary particles (also called monomers) that form at high altitudes and stick together to form aggregate particles as they sediment below 600 km to form the visible haze (West and Smith, 1991; Tomasko et al., 2008; Cours et al., 2011; Lavvas et al., 2013).

The detached haze layer was first identified in Voyager images at an altitude near 360 km near the equator and at somewhat lower altitudes at lower latitudes in the summer hemisphere (Rages and Pollack, 1983). It was seen near 500 km altitude in Cassini images when the spatial resolution became sufficient in 2005. It has undergone a remarkable drop in altitude, most rapidly near equinox in 2009 (West et al., 2011). The layer provides a marker to determine the center of Titan in images which in turn leads to an accurate measure of its altitude. At the time the images in Fig. 1 were obtained (2012-127, 6 May) the detached haze and its polar enhancement were measured to be 300 ± 10 km. Fig. 2 shows a circle at 300 km fitted to the detached haze in the CB3 image.

![Fig. 1. Mosaics of Titan against Saturn on 6 May, 2012 in the UV3 (343 nm) filter on the left and the CB3 (935) continuum filter on the right. Shadows of the rings appear mostly in the lower part of the image. The north angle (angle of the north-pointing end of the spin vector measured clockwise relative to the vertical) is –18.5° in both images.](image1)

![Fig. 2. A dashed (alternating black/white) circle at 300 km altitude shown superimposed in the 935-nm image of Titan in front of Saturn (image N1715033487) obtained on 2012-127 (6 May). The north angle is –18.5°.](image2)

![Fig. 3. Changes in the haze density at high southern latitudes are apparent in the two images shown here, both with Saturn in the background and both through the BL1 filter (455 nm). The image on the left (N1684691903) was obtained on day 141 (21 May) of 2011. The image on the right (N1715034531) was taken on day 127 (6 May) of 2012. The north angles are 0.0° and –18.5° respectively, for the left and right images. Arrows are aligned with the spin axis and point to the south pole.](image3)
The significance of the changes in particle density in the region of the south pole is obvious in before/after images under similar viewing geometry as shown in Fig. 3.

As seen in Fig. 3 the detached haze on day 141 (21 May) of 2011 appears to be uniform in its physical and optical thickness from mid-latitudes across the polar region. This configuration has been the norm from when it was first seen in Cassini images in 2005 until about 2012 where the right panel of Fig. 3 shows a physical and optical thickening at high southern latitudes.

We have examined the ISS image collection to see if the enhancement of the south polar haze can be traced to an earlier time (earlier than 2012-127 (6 May). Fig. 3 shows no enhancement as of day 141 (21 May) of 2011. We found several images taken on 2011-254 (11 September) that show a weak enhancement in scattered sunlight. These appear in Fig. 4. In the image on the left in Fig. 4 it is possible to see the detached haze (which has a blue color) from mid-latitudes all the way to the south pole. The intensity over the south pole is much stronger than at middle latitudes.

The haze is optically thin so there should not be any effect due to solar zenith angle except near the terminator where the solar flux becomes attenuated. This effect would diminish rather than enhance scattering by aerosols near the south pole relative to limb intensity at other latitudes, yet the opposite behavior is observed.

Limb and near-limb images of Titan throughout the period 2012-178 (26 June) to the present (Fig. 5) consistently show the altitude of the cloud to be at or near 300 km (with uncertainty

Fig. 4. These images from 2011-254 (11 September) show enhanced scattering by haze above the south pole (to the right in both images). The image on the left was constructed from three Wide Angle Camera (WAC) frames in blue, green and red filters. The image on the right was similarly constructed from three Narrow Angle Camera (NAC) images. The north polar angle is \(94.7^\circ\) for both. In the WAC image the intensity values were multiplied by the same factor for each of the three component images to bring forth the faint outer haze layer, but in doing so the brighter parts of the image were saturated and appear white.

Fig. 5. A dashed circle at altitude 300 km is superimposed on four images of Titan with the south polar cloud feature at or near the limb. The feature can be seen at all phase angles sampled here (see Table 1 for details). Arrows indicate the location of the cloud feature.
By coincidence (perhaps) the detached haze, seen as a local maximum in the $I/F$ profile along the limb, aligns with the cloud patch seen on the limb. The detached haze extends from the winter polar latitudes (55°N) all the way to the haze patch near the south pole. We were able to place a circle with radius 2875 km on top of the detached haze and thereby locate the center of Titan and show that the altitude of both the detached haze and south polar cloud patch was 300 km for images in Figs. 2 and 5. The uncertainty quoted above corresponds to one pixel. The uncertainty is based on previous experience as reported by West et al. (2011).

3. Near-nadir images

3.1. Color and photometry, including methane bands

On 2012-178 (26 June) from a vantage point above the high southern latitudes (Titan subspacecraft latitude –77.7°) the ISS cameras had an advantageous view of a newly formed cloud patch as shown in Fig. 6. At the time the image was obtained the subsolar latitude on Titan was 15°, so the cloud feature was illuminated by virtue of light that passed through the geometric terminator. That fact is apparent in the raw images where the cloud patch is difficult to perceive at the low light levels beyond the geometric terminator. The image shown in Fig. 6 has a shading correction applied to make the feature more visible.

Several color and morphologic features make this cloud distinct and different from the widespread haze seen previously at southern high latitudes and throughout the mid-latitudes. The cloud has a distinct boundary which appears from shading on the sunward side to be convex downward. The shading pattern suggests that the outer edge of the cloud is at a slightly different altitude than the interior, although this could be due to an enhanced particle density in this region rather than a shape effect. Within the interior the cloud has a mottled texture unlike the main haze. This suggests a clumpy density distribution with many local relative maxima and minima. Such a distribution is a natural configuration for a condensable material but not for an involatile haze. For this reason we refer to this feature as a cloud and not haze.

The color of the cloud is different from its immediate surroundings – more yellow than orange/brown and brighter. This color can also be seen in a faint boundary region about 40 km wide separating the main part of the cloud from the surrounding haze.

A more complete picture of photometric differences between the cloud and surrounding haze can be achieved by examination of ratios of the reflectivity under identical lighting and viewing geometry. Fig. 7 shows two regions marked by bright pixels chosen for comparison. Fig. 8 shows the ratio $I_{cloud}/I_{haze}$ for these regions as a function of effective wavelength in each of nine filters. At all wavelengths except near-UV the cloud is significantly more reflective than the haze.

The ratio has a local maximum in the green filter. The cloud is especially striking in the strong 889-nm methane filter. Fig. 9 shows a methane/continuum ratio image (889-nm MT3/935-nm CB3). The average MT3/CB3 ratio in the cloud patch is 0.60 while for the adjacent region it is 0.35. The higher ratio for the cloud patch is indicative of a high-altitude source region and is thus consistent with 300 km altitude or higher for the cloud patch as described next.
be the background intensity field. Then, we added the cloud as a lower boundary condition, with a radiance factor \( I/F_c \) at the top of the cloud, \( z_T \). We then integrated the source function from the cloud level to the top of the atmosphere to obtain the outgoing \( I/F \) in the presence of the cloud. Hence, the effect of the cloud is treated as an additive term that does not interfere with the background intensity field. We verified the validity of this approximation for SZA lower than 90° before performing the complete analysis.

To compute the background intensity field, we needed to define the atmospheric properties. The optical properties of the haze layer are taken from Tomasko et al. (2008), but we found that we need to scale the haze opacity by a factor \( F_h \sim 2 \) in order to fit the observations in the cloud-free zone that surrounds the cloud itself. Methane absorption is computed from the band model of Karkoschka and Tomasko (2010), and integrated over the instrument transmission (filter + detector sensitivity function) corresponding to the MT3 observation. The complete 3-D background intensity field is then simply recorded from the 3D-MC solutions. So, the radiative transfer equation can be re-integrated using any geometry. To compute the \( I/F_c \) that emerges from the top of the cloud, we modeled the cloud as a slab in a parallel plane model, and we related the upward \( I/F_c \) to its optical properties and to the geometry of the incident solar flux. The cloud is first defined by its optical thickness \( \tau_0 \). The phase function and single scattering albedo are calculated with Mie theory. We assume that the cloud is illuminated from below, with a solar flux attenuated by the transmission along its incident path through the atmosphere, treated in spherical geometry. The optical thickness along this incident path, \( \tau_0 \), is affected by several uncertainties: (1) the estimation of the scaling factor \( F_h \) on which it depends; (2) the assumption of spherical symmetry in the atmospheric properties to compute \( \tau_0 \) while in reality the properties vary with latitude; (3) errors introduced in the vertical structure of haze and methane absorption that are not well known. We estimate an uncertainty of at least 20% on the value of \( \tau_0 \). Then, with the SPSDISORT radiative transfer model (Mayer and Kylling, 2005), we compute the local radiative transfer solution inside the cloud and we determine the \( I/F_c \) that emerges from the top of the cloud as a function of the geometry and of the cloud properties.

Cassini VIMS (Visible and Infrared Mapping Spectrometer) observations demonstrate that the observed cloud is composed by HCN ice particles (de Kok et al., 2014). The cloud droplet radius is assumed to be as small as 1 µm, in relation with the amount of available material for condensation and the high settling speed at these altitudes. Therefore, in our calculations, we use as optical constants for the cloud particles, the real refractive index of the HCN ice while for the imaginary part, due to the lack of information, is left as a free parameter between \( 10^{-5} \) and \( 10^{-2} \). The effective size of particles is allowed to vary between 1 µm and 5 µm, whereas the effective variance is fixed at 0.2.

Fig. 10 shows, for a given pixel of the image, the variation of the simulated \( I/F_c \) of a cloud located at 310 km as a function of the cloud optical depth, for different values of the effective radius \( r_{\text{eff}} \) and of the imaginary part of the refractive index, \( k \). The black solid line represents the \( I/F_c \) value deduced from the observations, calculated as the difference between the observed \( I/F \) for the chosen pixel and the computed background \( I/F \) field, both divided by the transmission along the emerging line of sight. The cloud optical depths that match the observations are given, for each case, by the intersections between the observed \( I/F \) for the pixel chosen (black solid line) and the simulated \( I/F_c \). We find that observations may correspond to two different values of the cloud opacity, only one value or no value depending on the case. In this example, no solution can be found for \( k = 10^{-3} \), while for very low values of \( k \), we generally find two possible values of the cloud opacity.

**Fig. 8.** The ratio of intensity in the cloud divided by intensity in the haze at the same lighting and viewing geometry is shown as a function of effective wavelength for each of nine filters. * symbols are in methane absorption bands; X symbols are at continuum wavelengths.

**Fig. 9.** This view shows a ratio of \( I/F \) in the methane MT3 and CB3 filters using images N1719446402 and N1719445687, respectively.

### 4. Cloud optical properties

In order to retrieve the cloud properties, we analyzed the radiance factor \( I/F \) in the methane 890-nm (MT3) filter by using radiative transfer simulations. The cloud simulation requires the use of a three-dimensional Monte-Carlo radiative transfer model (Trân, 2005) in spherical geometry (referred hereafter as the 3D-MC) since the plane-parallel approximation breaks down for high solar zenith angle (SZA). However, because the cloud is not spherically symmetric but has a finite size and is located near the pole, we needed to evaluate separately its contribution to the intensity. We performed the calculation in two steps. We first determined the source function computed by the 3D-MC model in the spherically symmetric atmosphere including the effects of haze and methane absorption, but without the cloud. This is assumed to
We selected for radiative transfer analysis 146 pixels at the edge of the cloud. For each pixel, we computed \( I/F \) for cloud altitudes ranging from 300 km to 320 km in steps of 10 km, \( r_{\text{eff}} = 1 \mu m, 3 \mu m \) and 5 \( \mu m \) and using the same values of \( k \) as presented in Fig. 10. These cloud altitudes were chosen according to the results presented in Section 2. In particular, for a cloud top altitude of \( z_T = 320 \) km, no solution can be found for \( k = 10^{-5} \), for cloud droplet size \( r_{\text{eff}} = 1 \mu m, 3 \mu m \) and 5 \( \mu m \). In the case of \( z_T = 310 \) km and 300 km, no solutions are found for all pixels used in the analysis. In particular, for a cloud altitude of 310 km our model can fit only 40\% of the sampled pixels, whereas this percentage is reduced to 4\% for \( z_T = 300 \) km.

With an uncertainty of \( \pm 20\% \) assumed on the value of \( \tau_0 \), we find that observations are fitted for 100\% and 75\% of the pixels used for cloud altitudes of 310 km and 300 km respectively. These results show that for any cloud height and \( r_{\text{eff}} \), no solution can be found for \( k \) values larger than about \( 10^{-4} \), while for very low values of \( k \), we generally find two possible values of the cloud opacity (\( \tau_{c1} \) and \( \tau_{c2} \)). Table 2 shows the mean values of \( \tau_{c1} \) and \( \tau_{c2} \) with their standard-deviation, for each cloud altitude, effective radius, and imaginary refractive index smaller than \( 5 \times 10^{-4} \). We find that in all cases, we have two solutions of the cloud optical depth (\( \tau_{c1} \) and \( \tau_{c2} \)) that are equally likely from our photometric considerations. However, we know from recent observations made with VIMS that this cloud is transparent enough to see through it to surface features (Le Mouélic et al., 2012). Hence, only the optical depth of the cloud given by the first solution of the model (\( \tau_{c1} \)) is really consistent with observations. We then find that the optical thickness in the vertical direction is in a range between 0.6 and 2.6 and the imaginary refractive index, \( k \), should be smaller than about \( 5 \times 10^{-4} \).

### Table 2

Cloud optical depths for a variety of particle refractive indices, radii, and altitudes.

<table>
<thead>
<tr>
<th>( k )</th>
<th>( r = 1 \mu m )</th>
<th>( r = 3 \mu m )</th>
<th>( r = 5 \mu m )</th>
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<td>( \tau_{c1} )</td>
<td>( \tau_{c2} )</td>
<td>( \tau_{c1} )</td>
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<td>1 \times 10^{-4}</td>
<td>1.6 ± 0.6</td>
<td>6.9 ± 1.8</td>
<td>1.6 ± 0.6</td>
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<tr>
<td>5 \times 10^{-4}</td>
<td>1.5 ± 0.6</td>
<td>6.9 ± 1.8</td>
<td>1.7 ± 0.7</td>
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<td>1 \times 10^{-4}</td>
<td>1.1 ± 0.4</td>
<td>7.8 ± 1.2</td>
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<td>1 \times 10^{-4}</td>
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<td>9.1 ± 0.8</td>
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<td>5 \times 10^{-4}</td>
<td>0.8 ± 0.2</td>
<td>9.1 ± 0.8</td>
<td>1.6 ± 0.6</td>
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</table>

Fig. 10. Simulation of \( I/F \) for an altitude of 310 km and for different values of \( k \) and \( r_{\text{eff}} \). The black solid line represents the observed \( I/F \) value for the chosen pixel and the dashed lines the error.
was then used to remove the large-scale illumination gradient and isolate small-scale features. The visible limb was used to adjust the predicted pointing from C-kernels and define a latitude–longitude grid. For these polar images we performed tracking in the original image coordinate plane, using tracking boxes of 20 × 20 pixels, roughly 2.77 km on a side. In addition to the quality control criteria discussed by Del Genio and Barbara (2012), we also excluded wind vectors whose tracking target boxes have a standard deviation of pixel range < 4% of the dynamic range of the image. This helps exclude some very low contrast areas outside the collar for which tracking is not reliable. The derived displacements were then converted to zonal and meridional winds using the navigation and image time separation.

Achterberg et al. (2008a) found that isotherms in Titan’s stratosphere are offset from the poles by 4.1° in a sense that suggests that the stratospheric circulation may adjust seasonally to produce a circulation spin equator that is slightly offset from that of the solid body so as to optimize the effect of near-equatorial solar heating on upward angular momentum transport. The equinox seasons may be the times when this solar bias of the axis is most strongly reinforced. Assuming approximate gradient wind balance, a similar offset should be expected in the wind field. An offset manifests itself as a wavenumber one departure of the winds from their zonal mean values, so the optimum offset can be estimated by minimizing the wind variance. We calculated the variance in the zonal and meridional winds at each latitude for various offsets of the pole relative to the navigation best estimate. This was done in two stages. The first stage used coarse resolution (32 pixel increments) searches in both directions, with the minimum variance being found at a location displaced 64 pixels from the navigation-based pole (Fig. 12, upper panels). A second fine resolution (8 pixel increments) search (Fig. 12, lower panels) produces a best estimate offset that places the circulation pole at 85.5°S, 263.7°W in the original navigation-based coordinate system. The magnitude of the offset is consistent with that found by Achterberg et al. (2008a) and the sense is such as to better align the spin equator with the Sun (though not completely, since the subsolar latitude was 15° at the time of these observations).

The resulting zonal and meridional wind profiles using the best estimate pole offset are shown in Fig. 13. Wind vectors are also superimposed on the image in Fig. 11. The zonal wind profile peaks at the latitudes of the bright polar collar and decreases poleward from there. Tracking is difficult within and outside the collar because of the diffuse contrast and lack of small-scale features, and the tracking algorithm fails at times despite the quality criteria applied, producing some vectors with small or retrograde wind
speeds. Visual inspection of the features responsible for a number of these vectors confirms that they are artifacts rather than successful tracking estimates. Closer to the pole, these anomalous wind vectors are absent. There also appears to be real variance of \(\sim 10-20 \text{ m s}^{-1}\) in the wind field, some of it probably due to the wavenumber 2 behavior of the collar. The solid line shows the mean zonal wind profile, which is clearly biased \(20-30 \text{ m s}^{-1}\) low by the artifacts. The median performs somewhat better but is still shifted \(\sim 10 \text{ m s}^{-1}\) low. We therefore binned the zonal winds at each latitude into \(1 \text{ m s}^{-1}\) increments and the meridional winds into \(0.1 \text{ m s}^{-1}\) increments. The dashed line shows the mode value of these distributions at each latitude, which is probably a better estimate of the true average.

Table 3 gives the mean and mode wind values as a function of latitude. The zonal wind peaks near the latitude of the bright collar and decreases inside. The mode value at \(82^\circ\)S is consistent with the estimated \(87-89 \text{ m s}^{-1}\) propagation speed of the feature rotation derived earlier from the hourly images, suggesting that the cloud patch is quasi-stationary within the wind field. The zonal wind speed produced by an angular momentum-conserving displacement of an air parcel from rest at the equator, given by \(u_m = -\Omega \sin^2 \theta / (\cos \theta)\), where \(\Omega\) is Titan’s angular rotation frequency, \(a\) is Titan’s radius at \(300 \text{ km}\) altitude, and \(\theta\) is latitude, is \(92 \text{ m s}^{-1}\). This suggests that eddy effects on the wind are small at this latitude.

Meridional winds are also shown in Fig. 13. These are unsurprisingly near zero, although non-negligible variance about the mean exists, some of which is likely to be real. Table 3 shows no signal in the mean meridional wind, but the mode value is consistently negative from the collar to the pole. General circulation models suggest that Titan’s stratospheric circulation spends much of the year in single-cell solstitial configurations, with rising motion at the summer pole and sinking at the winter pole, to balance the radiative heating and cooling, respectively, at the two poles (e.g., Lebonnois et al., 2012). Near the equinoxes a two-cell circulation with rising motion at the equator is predicted to exist. The observations analyzed here occur between Titan’s vernal equinox and northern summer solstice. The implied meridional circulation at \(300 \text{ km}\) in the Lebonnois et al. (2012) simulations is poleward in the southern hemisphere in both the equinox and solstice configurations, so we cannot tell from these observations whether the expected seasonal shift had occurred by this time. However, the fact that an angular momentum-conserving poleward flow is consistent with the observed zonal winds, combined with the inferred negative (poleward) mode meridional wind we derive, appear at least qualitatively consistent with expectations for either seasonal configuration. The magnitude of the mode meridional wind is a factor of \(\sim 2-5\) stronger than that expected from scaling arguments using an estimate of the radiative heating/cooling combined with the continuity equation (Achterberg et al., 2008b). On the other hand, Achterberg et al.’s vertical velocity estimate of \(\sim 0.5 \text{ mm s}^{-1}\) is at the low end of the ranges derived by Teanby et al. (2012) from their retrievals of changes in temperature \((0.5-2.0 \text{ mm s}^{-1})\) and trace gas abundances \((0.8-2.3 \text{ mm}^{-1})\), suggesting that our tracked mode meridional wind speeds may be reasonable.

Formally, an individual wind vector can probably be tracked to an accuracy no better than \(\sim 5 \text{ m s}^{-1}\) given the pixel resolution and image time separation, but the accuracy of the mode value at a given latitude may be higher, even more so given that the

### Table 3

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<th>((v)) (m/s)</th>
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Fig. 13. Zonal (left) and meridional (right) wind profiles through the south pole cloud patch. The solid curves indicate the zonal mean, while the dashed curves are the profiles of mode wind values.
sign is consistent at all latitudes. On the other hand, our derived pole offset is an inference, however plausible, with its own uncertainty, so the estimated meridional wind should be viewed circumspectly.

6. Discussion and conclusions

The morphology of the small-scale haze structure within the collar is reminiscent of open-celled convection that occurs at low altitudes on Earth (Fig. 14). This morphology is preferentially observed in two distinct meteorological conditions: Off the east coasts of continents in cold-air outbreaks behind cold fronts associated with synoptic midlatitude storms, and in the middle of extensive stratocumulus decks that occur frequently in the sub-tropics off the west coasts of continents (Mühlbauer et al., 2014). Both situations have certain meteorological characteristics in common: (1) convection occurs as cooler air is advected over a warmer surface,destabilizing the lapse rate and allowing rising turbulent boundary layer air parcels to become buoyant; (2) the convection remains shallow because the prevailing large-scale vertical velocity in the free troposphere above the clouds is downward, with the resulting downward advection of heat along a positive potential temperature gradient creating an inversion layer that suppresses vertical development of the convection.

The cellular morphology in the interior of the Titan south pole cloud patch is probably more similar in nature to the subtropical stratocumulus decks seen in Fig. 14, but the circumstances are somewhat different. In terrestrial stratocumulus clouds, the turbulence that maintains the cloud layer by mixing evaporated water upward from the ocean surface is driven by longwave radiative cooling from the cloud top. As the cloudy air mass is advected southward over warmer sea surface temperatures, the cloud layer often becomes decoupled from its surface sources of moisture, and once shallow convection breaks out in the cloud layer, the subsiding air outside the convective cloud dries out the stratocumulus deck and begins to clear out the cloud, as seen in Fig. 14. On Titan, the emergence of the south polar cloud during southern autumn suggests that it is a response to slow seasonally varying radiative cooling. By this time the polar atmosphere is more or less in darkness and is cooling radiatively to space, and the decrease of the radiative relaxation time with height implies that higher altitudes may respond more quickly to the seasonal insolation change than lower altitudes. The condensable species that forms the cloud feature most likely is from a source above the cloud and its concentration determined by the general circulation. The resulting radiative flux divergence at the cloud top might be the source of the destabilization that creates the cellular morphology of the clouds. The fact that by this time of the Titan year the mean circulation is probably already producing downward motion at the south pole at 300 km altitude implies that vertical development of any convective motion is suppressed, just as for situations that lead to open cells on Earth.

The feature contrast may be due to something as simple as particles preferentially remaining aloft against sedimentation in the convective upwelling locations while sedimentation is enhanced in the subsiding regions outside the convective updrafts. It could also be the result of a condensation–evaporation cycle associated with small-scale upwelling and downwelling motions as occurs in open cell water cloud on Earth. For Titan, de Kok et al. (2014) have reported the detection of HCN ice in the south polar region at altitudes consistent with the clouds seen by ISS. De Kok et al. show that south polar temperatures near 300 km altitude were already much colder than those predicted by general circulation models by September 2011 and had cooled ~10 K further by February 2012, to within ~40–50 K of the HCN saturation temperature. Temperature retrievals at 300 km do not exist after that, but retrievals up to ~225 km altitude in October 2013 show this region to have cooled to the HCN saturation point, so HCN ice is at least a plausible candidate for the ISS south polar cloud feature. In keeping with this idea, the distinct color of the polar cloud patch relative to its surroundings suggests a different history for these particles than the more ubiquitous Titan reddish haze particles that primarily sediment out from their production region aloft.

Some further surprises may be in store. Two coincidences related to the south polar cloud patch emerge from the collection of observations made with Cassini instruments. One of these is the coincidence in the altitude of the haze patch (300 km) and the altitude of the detached haze in 2012–2013 extending to equatorial latitudes or higher. The circle at 300 km altitude used to determine the altitude of the haze patch as seen in Figs. 2 and 5 was centered on Titan by matching with the detached haze local I/F maximum over a large range of latitudes. Do the conditions responsible for the detached haze somehow serve as a trigger for the formation of the south polar haze patch?

We remarked in the introductory section of this paper that Jennings et al. reported the formation of a south polar feature with a spectral signature at 220 cm⁻¹ coincident with the formation of the haze patch seen at visible wavelengths. Jennings et al. point out that such a feature was seen in the norther polar region prior to formation in the south and that its altitude was determined to be much lower than 300 km. The temporal coincidence of formation of these to features in the south suggested that the same cloud was observed at by all three Cassini instruments (CIRS, VIMS, ISS). However, the identification of HCN ice at 300 km altitude by de Kok et al. (2014) and the lower altitude for the northern 220 cm⁻¹ feature suggest that the two are different features with different compositions that happened to form in the south at about the same time. Perhaps further scrutiny of the data will reveal if the 220 cm⁻¹ feature is too deep to be seen by VIMS and if the cloud (presumably HCN) at 300 km can be seen by CIRS when better spatial resolution becomes available.

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