The origin and evolution of Saturn's 2011–2012 stratospheric vortex

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Abstract

The planet-encircling springtime storm in Saturn's troposphere (December 2010–July 2011, Fletcher, L.N. et al. [2011]. Science 332, 1413–1414; Sánchez-Lavega, A. et al. [2011]. Nature 475, 71–74; Fischer, G. et al. [2011]. Nature 475, 75–77) produced dramatic perturbations to stratospheric temperatures, winds and composition at mbar pressures that persisted long after the tropospheric disturbance had abated. Thermal infrared (IR) spectroscopy from the Cassini Composite Infrared Spectrometer (CIRS), supported by ground-based IR imaging from the VISIR instrument on the Very Large Telescope and the MIRSI instrument on NASA's IRTF, is used to track the evolution of a large, hot stratospheric anticyclone between January 2011 and March 2012. The evolutionary sequence can be divided into three phases: (I) the formation and intensification of two distinct warm airmasses near 0.5 mbar between 25 and 35°N (B1 and B2) between January–April 2011, moving westward with different zonal velocities, B1 residing directly above the convective tropospheric storm head; (II) the merging of the warm airmasses to form the large single 'stratospheric beacon' near 40°N (B0) between April and June 2011, disassociated from the storm head and at a higher pressure (2 mbar) than the original beacons, a downward shift of 1.4 scale heights (approximately 85 km) post-merger; and (III) the mature phase characterised by slow cooling (0.11 ± 0.01 K/day) and longitudinal shrinkage of the anticyclone since July 2011. Peak temperatures of 221.6 ± 1.4 K at 2 mbar were measured on May 5th 2011 immediately after the merger, some 80 K warmer than the quiescent surroundings. From July 2011 to the time of writing, B0 remained as a long-lived stable stratospheric phenomenon at 2 mbar, moving west with a near-constant velocity of 2.70 ± 0.04 deg/day (–24.5 ± 0.4 m/s at 40°N relative to System III longitudes). No perturbations to visible clouds and hazes were detected during this period.

With no direct tracers of motion in the stratosphere, we use thermal windshear calculations to estimate clockwise peripheral velocities of 200–400 m/s at 2 mbar around B0. The peripheral velocities of the two original airmasses were smaller (70–140 m/s). In August 2011, the size of the vortex as defined by the peripheral collar was 65° longitude (50,000 km in diameter) and 25° latitude. Stratospheric acetylene (C2H2) was uniformly enhanced by a factor of three within the vortex, whereas ethane (C2H6) remained unaffected. The passage of B0 generated a new band of warm stratospheric emission at 0.5 mbar at its northern edge, and there are hints of warm stratospheric structures associated with the beacons at higher altitudes (p < 0.1 mbar) than can be reliably observed by CIRS nadir spectroscopy. Analysis of the zonal windshear suggests that Rossby wave perturbations from the convective storm could have propagated vertically into the stratosphere at this point in Saturn's seasonal cycle, one possible source of energy for the formation of these stratospheric anticyclones.

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Atmospheres, Structure
1. Introduction

The outbreak and evolution of a spectacular springtime storm in Saturn’s northern mid-latitudes in 2010–2011 (Fletcher et al., 2011; Sánchez-Lavega et al., 2011; Fischer et al., 2011) captured the imaginations of amateurs and professionals alike, spurring an international observing campaign to support the orbital high-resolution observations from the Cassini spacecraft. Infrared observations by the Cassini Composite Infrared Spectrometer (CIRS, Flasar et al., 2004) and the Very Large Telescope (VLT) determined the thermal structure of a saturnian storm for the first time, revealing that the convective storm cells seen in visible imaging produced unexpectedly large perturbations to stratospheric temperatures near 20–40°N (all latitudes in this article are given as planetographic, and all longitudes as System III west Seidelmann et al., 2007), hundreds of kilometres above the roiling cloud decks of the troposphere. Those stratospheric anomalies, referred to as ‘beacons’ of emission, were expected to cool as the storm abated, but instead their evolution through 2011 and 2012 have revealed an entirely new phenomenon in a giant planet stratosphere: the formation and evolution of a giant stratospheric vortex. This article traces the temperatures, winds and composition of Saturn’s stratosphere as the vortex evolved in 2011 and 2012.

Historical records in reflected sunlight over the past 130 years show that Saturn’s planetary-scale disturbances exhibit an episodic behaviour and usually occur after summer solstice (Sánchez-Lavega et al., 1991), with kronocentric solar longitudes (i.e., Saturn-centric, \( \lambda_s \)) in the range from 110° to 167°, measured from the spring equinox at \( \lambda_s = 0° \) (August 15th, 2009). However, this current storm started on December 5th 2010 (\( \lambda_s = 16° \)), the active convection stage persisted until around July 2011 (\( \lambda_s = 23° \)), and the stratospheric beacon remains present at the time of writing (\( \lambda_s = 32° \)). This is 9 Earth-years (more than a saturnian season) earlier than would be expected from the normal summertime storm occurrence, making this our first example of a planetary-scale springtime storm. There is evidence that spring is an important time, dynamically, on other planets too, as revealed by enhanced cloud activity during Uranus’ 2007 equinox (e.g., Sromovsky et al., 2009), although the mechanisms driving this seasonal activity are poorly understood.

The cloud-top evolution of the tropospheric disturbance has been well-documented in reflected sunlight from both the Cassini spacecraft (Sayanagi et al., 2012) and ground-based observers (Sánchez-Lavega et al., 2012), and shows similarities with previous ‘Great White Storm’ events on Saturn (Sánchez-Lavega, 1982; Sánchez-Lavega et al., 1991). The first visible observations of the storm coincided with the detection of abundant electrical activity by the Radio and Plasma Wave Science instrument onboard Cassini (Fischer et al., 2011). The powerful upwelling of moist air in multiple storm cells in the convecting area was accompanied by near-continuous lightning discharges driven by Saturn’s internal heat (peaking at ten Saturn electrostatic discharges, or SEDs, per second, Fischer et al., 2011). However, the unprecedented resources provided by the infrared instrumentation on Cassini and the high spatial resolution afforded by the 8-m class of ground-based observatories provided our first observations of the thermal structure of a saturnian storm (Fletcher et al., 2011).

Thermal imaging of the storm showed that a cold, compact (4000 × 5500 km, 800 km) anticyclonic vortex had formed in the upper troposphere near 41°N planetographic latitude to the east of the storm head within the first few days of the storm onset. This large, cold tropospheric oval in Fig. 1 of Fletcher et al. (2011) appeared a dark bluish colour in visible imaging and was surrounded by a collar of whiter clouds. The oval propagated westward with the retrograde jet at 39°N, albeit slower than the velocity of the storm head (Sánchez-Lavega et al., 2011). The tropospheric vortex marked a distinct boundary between the storm head to the west and the easterly tail, rather like Jupiter’s Great Red Spot perturbing jet streams around its periphery to create the turbulent wake of Jupiter’s South Equatorial Belt (SEB).

VLT and Cassini imaging, sensitive to stratospheric emission from methane and ethane (7.9, 8.6 and 12.3 μm), demonstrated substantial perturbations to the stably-stratified upper atmosphere to pressures as low as 0.5 mbar (Fletcher et al., 2011). The compact tropospheric vortex was flanked at high altitude to the east and west by the diffuse warm airmasses known as stratospheric beacons. In this paper we extend the observations to track the surprising evolution of the beacons after their initial discovery. In Section 2 we discuss the Cassini and ground-based instruments used to track Saturn’s stratospheric emission from January 2011 to March 2012, plus our analysis and modelling techniques. The evolution of the beacons is described in Section 3, before we determine the atmospheric temperatures, winds, stability and hydrocarbon composition associated with the beacons in Section 4. We discuss mechanisms and numerical simulations for the origins of the stratospheric beacons in Section 5 before highlighting the key conclusions of this infrared study in Section 6.

2. Sources of data

The evolution of Saturn’s stratospheric thermal-infrared (IR) emission was monitored throughout 2011 and 2012 using three different resources: spectroscopic mapping from the Composite Infrared Spectrometer (CIRS, Flasar et al., 2004) on the Cassini orbiter (observation details in Table 1); high spatial resolution imaging in the N band (8–14 μm) from the ESO (European Southern Observatory) Very Large Telescope (VLT) mid-IR camera/spectrometer (VISIR, Lagage et al. (2004) observation details in Table 2); and moderate spatial resolution N-band imaging from NASA’s Infrared Telescope Facility (IRTF) mid-IR spectrometer and imager (MIRSI, Deutsch et al., 2003 observation details in Table 3). We describe the data acquisition and reduction in the following sections, before discussing the observations in Section 3.

2.1. Cassini/CIRS spectroscopy

Cassini/CIRS (see Flasar et al. (2004) for a complete description) comprises two interferometers fed by a shared telescope and fore-optics. The polarising Martin–Puplett far-IR interferometer (10–600 cm⁻¹, known as focal plane 1, or FP1) will not be used in the present stratospheric analysis, as its sensitivity is primarily tropospheric. Instead, we use the mid-IR Michelson interferometer, which features two arrays of \( 1 \times 10 \) HgCdTe detectors (focal planes 3 and 4, or FP3, 600–1100 cm⁻¹ and FP4, 1100–1500 cm⁻¹) with an instantaneous field of view (IFOV) of 0.27 × 0.27 mrad. Observation uses pairs of pixels coadded as a single 0.27 × 0.54 mrad pixel. CIRS has a programmable spectral resolution from 0.5 to 15.5 cm⁻¹, with the highest spectral resolutions used for compositional studies and the lowest resolution used for mapping temperature perturbations.

Four different observational templates were used to study the beacon evolution (Table 1). The storm was mapped at the highest spatial resolution of 0.4° latitude (equivalent to 310 km at 40°N) on August 21st 2011 and January 27th 2012 using the 15.5 cm⁻¹ spectral resolution setting used for mapping (known as a FIRMAP), where the fields of view are repeatedly scanned along the central meridian as Saturn’s 10-h rotation sweeps out a longitude circle. The majority of observations used a ‘sit-and-stare’ technique at a particular latitude at 0.5 or 2.5 cm⁻¹ spectral resolution (known
as COMPSITs and MIRMAPs, respectively), aligning the arrays north-south and allowing Saturn’s rotation to sweep the arrays across all longitudes. These observations were taken at a greater distance from Saturn, yielding spatial resolutions of 1–2° longitude (800–1600 km at 40°N) and 2–4° latitude in swaths 10–20° latitude wide. Those observations that captured part of the stratospheric disturbance are recorded in Table 1. Finally, MIRMAPs use the 2.5 cm⁻¹ spectral resolution setting, but in this case the
focal planes are stepped from north to south (‘shift and stare’),
dwelling on a particular latitude for approximately 1.5 h (50°
of longitude). All spectra were extracted from the most recent Cas-
sini/CIRS calibrated database (version 3.2, which uses deep space
interferograms within 18 h of the target interferograms for radio-
metric calibration), and were coadded to improve the signal to
noise as necessary in Section 4. As the number of target interfero-
grams in each observation exceeds the number of cold deep space
 calibration spectra, we find that the noise equivalent spectral radi-
ance (NESR), and hence the retrieval uncertainty for each bin, is
dominated by the measurement uncertainties on the deep space
spectra.

2.2. VLT/VISIR imaging

The spectroscopic observations from Saturn orbit were supple-
mented by ground-based VLT thermal-IR filtered imaging to pro-
vide contextual information for the stratospheric disturbance
between January and July 2011 when the planet was visible (Table
2). VISIR observations (Lagage et al., 2004) were obtained in two
programs using the Melipal (UT3) telescope at Cerro Paranal in
Chile: a regular program 386.C-0096 (January–May 2011) and
Director’s Discretionary program 287.C-0096 (June–July 2011).
The goal for each run was to obtain images of Saturn at eight dis-
crete wavelengths – 17.6, 18.7 and 19.5 μm to constrain tropo-
pheric temperatures from the H₂–He collisionally-induced
continuum; 9.0 and 10.7 μm for tropospheric PH₃; and 7.9, 8.6
and 12.3 μm for stratospheric temperatures via emission from
CH₄, CH₃D and C₂H₆, respectively. It is these latter three filters that
have been used in this study.

Quantitative comparison of filtered images is challenging – see-
ing conditions and the water vapour humidity varied between each
run, and the range to Saturn varied between 8.6 and 9.9 AU during
the 2011 campaign. This meant that the diffraction-limited spatial
resolution using the 8.1-m diameter primary mirror varied be-
tween 1530 and 1760 km for 7.9 μm (0.25λ), and 2380–2740 km
for 12.3 μm (0.38λ) on Saturn’s disc. In practice, the seeing could
degradation to 1.0’ during an observation, equivalent to 6240–
7180 km on Saturn’s disc, depending on the Earth to Saturn dis-
tance. However, these ground-based images proved vital in tracing
the motion and size of the beacons in Section 3.

Two or more images were obtained in each filter, dithered on
the array to allow removal of bad pixels. Images on each date were
selected based on their display of the beacon, and are shown in
Fig. 1. The ESO data pipeline was used for initial reduction and
bad-pixel removal via its front-end interface, GogoMo (version
2.3.0). Images were geometrically registered and cylindrically
reprojected using the techniques described in Fletcher et al.
(2009c). Images in each filter were radiometrically calibrated by
scaling the flux to match a Cassini/CIRS FIRMAP (15.5 cm⁻¹) ob-
ervation of the southern hemisphere on October 18th 2010 (i.e.,
avoiding the northern perturbations). The CIRS data were con-
volved with the VISIR filter functions prior to the scaling.

2.3. IRTF/MIRSI imaging

In addition to VLT/VISIR observations, we undertook a campaign
of MIRSI (Deutsch et al., 2003) mid-IR imaging using the 3-m pri-
mary mirror of the IRTF (Table 3). These were subdivided into six
discrete observing runs between March and September 2011, and
each observing run featured imaging in a number of discrete filters,
similar to those given for VLT/VISIR. Only the 7.9 and 12.3-μm fil-
ters are considered in this study. During the course of the 2011
observations, Saturn varied from 8.6 to 10.4 AU from Earth, imply-
ning that the diffraction-limited spatial resolution varied from 4130
to 5000 km at 7.9 μm (0.66′′) and 6430–7780 km at 12.3 μm (1.0′′).
The data acquisition and reduction procedures are described in
Fletcher et al. (2009c): chopping and nodding were used to elimi-
nate the background sky emission and thermal interference; sky
subtraction and flat-fielding were applied to remove variations in
the background across the detector; and five images (dithered
around the array to eliminate corrupted pixels) were coadded to
produce the final MIRSI images. These MIRSI images were used
to monitor the evolution of the beacons in Section 3.

2.4. Spectral modelling

Inspection of the raw data can tell us a great deal (e.g., mapping
the spatial extent and position of the stratospheric airmasses in Sec-
ton 3), but a better understanding of the atmospheric effects of the
2010 storm can be gained by modelling the CIRS spectra. We employ
a radiative transfer and optimal estimation retrieval code (Nemesis,
Irwin et al., 2008) to determine the vertical temperature structure
from the CH₄ and H₂ emissions, and the molecular abundances of
hydrocarbons from their emission features. The retrieval model has
been fully described elsewhere (Irwin et al., 2008; Fletcher et al.,
2010), but we review the important details here.

Saturn’s reference atmospheric structure was defined on 120
pressure levels equally spaced in log(p) between 1 pbar and
10 bar. An a priori temperature profile was estimated as a mean
of Cassini/CIRS T(p) profiles between ±45° latitude from Cassini’s
prime mission using nadir data from Fletcher et al. (2010) (sensi-
tive to 1–800 mbar) and limb results from Guerlet et al. (2009)
(sensitive to 1 pbar to 20 mbar). The helium abundance He/H₂
was set to 0.135 (Conrath and Gautier, 2000). The PH₃ profile
was initially set to the CIRS-derived mole fraction of 6.4 ppm at

<table>
<thead>
<tr>
<th>Date</th>
<th>Start time (UTC)</th>
<th>Stop time (UTC)</th>
<th>Program ID</th>
<th>Comments</th>
</tr>
</thead>
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<tr>
<td>2011-03-26</td>
<td>13:53</td>
<td>15:42</td>
<td>011</td>
<td>B1 on the rising limb</td>
</tr>
<tr>
<td>2011-04-26</td>
<td>05:56</td>
<td>09:17</td>
<td>006</td>
<td>Tail of B1 emission, rising on western limb</td>
</tr>
<tr>
<td>2011-04-27</td>
<td>05:14</td>
<td>09:05</td>
<td>006</td>
<td>B1 rising on the limb</td>
</tr>
<tr>
<td>2011-05-23</td>
<td>08:27</td>
<td>10:44</td>
<td>(Engineering time)</td>
<td>B1 on central meridian</td>
</tr>
<tr>
<td>2011-07-24</td>
<td>02:21</td>
<td>03:18</td>
<td>010</td>
<td>Beacon tail region</td>
</tr>
<tr>
<td>2011-07-27</td>
<td>02:17</td>
<td>03:28</td>
<td>010</td>
<td>B0 near centre disc in poor seeing</td>
</tr>
<tr>
<td>2011-08-02</td>
<td>02:22</td>
<td>04:20</td>
<td>(Engineering time)</td>
<td>Tail region of B0</td>
</tr>
<tr>
<td>2011-08-03</td>
<td>02:03</td>
<td>04:20</td>
<td>(Engineering time)</td>
<td>B0 on setting limb</td>
</tr>
<tr>
<td>2011-08-04</td>
<td>02:42</td>
<td>04:34</td>
<td>(Engineering time)</td>
<td>B0 rising to central meridian</td>
</tr>
<tr>
<td>2011-08-05</td>
<td>02:18</td>
<td>04:19</td>
<td>(Engineering time)</td>
<td>B0 from centre to setting limb</td>
</tr>
<tr>
<td>2011-08-31</td>
<td>01:05</td>
<td>03:06</td>
<td>027</td>
<td>B0 on setting limb</td>
</tr>
<tr>
<td>2011-09-01</td>
<td>00:05</td>
<td>03:56</td>
<td>027</td>
<td>Tail region of B0, final MIRSI observation</td>
</tr>
</tbody>
</table>
\( p > 0.55 \text{ bar} \), decreasing due to photolysis at lower pressures with a fractional scale height of 0.27 (the ratio of the \( \text{PH}_3 \) scale height to the scale height of the bulk atmosphere, Fletcher et al., 2009a). The vertical distribution of \( \text{NH}_3 \) had a reference mole fraction of 100 ppm below 1 bar, decreasing with altitude following a saturated vapour pressure profile (\( p > 0.3 \text{ bar} \)) and a linear extrapolation to low pressures to represent photolysis (\( p < 0.3 \text{ bar} \)). The \( \text{CH}_4 \) mole fraction was set to a deep value of \( 4.7 \times 10^{-3} \) (Fletcher et al., 2009b) and drops with altitude due to both diffusive processes and photochemical destruction towards the homopause, as reviewed by Moses et al. (2000). Our \textit{a priori} hydrocarbon profiles were taken as the low-latitude mean (between \( \pm 30\degree \) latitude) of profiles published by Guerlet et al. (2009) using CIRS limb spectroscopy, which were themselves based on photochemical modelling from Moses and Greathouse (2005). This reference atmosphere is available from the principal author on request.

The sources of spectroscopic linedata for these gases are presented in Table 4. These data were used to generate \( k \)-distributions (ranking absorption coefficients, \( k \), according to their frequency distribution, Irwin et al., 2008) using evenly-sampled wavenumber...
It is important to note the family of potential solutions to this ill-posed inverse problem that the presence of such an unusually warm feature in Saturn's stratosphere alters the spectral sensitivity to temperature changes, highlighting the increased spectral contribution from stratospheric temperatures in the latter case. For all CIRS observations, we first fit the 600–680 cm\(^{-1}\) and 1100–1370 cm\(^{-1}\) spectra (focal planes 3 and 4) simultaneously to determine a continuous \(T(p)\) profile and the fractional scale height of PH\(_3\) (i.e., fixing the deep PH\(_3\) below 550 mbar and letting its abundance vary throughout the photolysis region). These two spectral regions sense 60–230 mbar in the troposphere (30–190 mbar when the beacon is present) and 0.6–5.7 mbar in the stratosphere (1.1–5.6 mbar when the beacon is present). The ranges quoted are the FWHM of the contribution functions averaged over the respective band. \(A_\text{priori}\) temperatures at \(p > 230\) mbar are taken from a mean of the Cassini prime mission far-IR (focal plane 1) study (sensitive to 100–800 mbar, Fletcher et al., 2010).

Fig. 2. Contribution functions \((dR/dT, \text{the rate of change of radiance } R \text{ with temperature } T)\) calculated for both the reference atmosphere (a) and for the high-temperature conditions found within the stratospheric vortex (b). Spectra were calculated at 2.5 cm\(^{-1}\) resolution characteristic of the CIRS MIRMAP datasets. The legend shows that the same logarithmic scale is used for both panels, and highlights the increased contribution of stratospheric temperatures when the beacon is present, and the subtle changes in altitude sensitivity described in the main text. VISIR filter wavelengths of 7.7, 8.6 and 12.3 \(\mu m\) are shown for reference.

## Table 4
Sources of spectroscopic linedata. Exponents for temperature dependence \(T^n\) given in the final column.

<table>
<thead>
<tr>
<th>Gas</th>
<th>Line intensities</th>
<th>Broadening half width</th>
<th>Temperature dependence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Collision-induced absorption (CIA)</td>
<td>H(_2)-H(_2) opacities from Orton et al. (2007), plus additional H(_2)-He, H(_2)-CH(_4) and CH(_2)-CH(_4) opacities from Borysow et al. (1988) and Borysow and Frommhold (1986, 1987), respectively.</td>
<td>H(_2) broadened using a half-width of 0.059 cm(^{-1}) atm(^{-1}) at 296 K</td>
<td>(n = 0.44) (Margolis, 1993)</td>
</tr>
<tr>
<td>CH(_4), CH(_3)D</td>
<td>Brown et al. (2003)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C(_2)H(_6)</td>
<td>Vander Auwera et al. (2007)</td>
<td>0.11 cm(^{-1}) atm(^{-1}) at 296 K (Blass et al., 1987)</td>
<td>(n = 0.94) (Halsey et al., 1988)</td>
</tr>
<tr>
<td>C(_2)H(_2)</td>
<td>GEISA 2003 (Jacquinet-Husson et al., 2005)</td>
<td>Fits to data in Varanasi (1992)</td>
<td>(n = 0.70–0.01) (J) (is the rotational quantum number) (Salem et al., 2004)</td>
</tr>
<tr>
<td>PH(_3)</td>
<td>Kleiner et al. (2003)</td>
<td>(\gamma_a = 0.1078 – 0.0014) cm(^{-1}) atm(^{-1}) and (\gamma_m = 0.0618 – 0.0012) cm(^{-1}) atm(^{-1}) (Levy et al., 1993; Bouanich et al., 2004)</td>
<td>(n = 0.0618)</td>
</tr>
</tbody>
</table>

The contribution functions calculated in the nadir for the C\(_2\)H\(_2\) \(\nu_5\) band show that the peak sensitivity is at 2.5 mbar (unchanged when the beacon is present), with a FWHM range from 0.3 to 10.7 mbar. Likewise for the C\(_2\)H\(_6\) \(\nu_3\) band, the sensitivity peaks at 2.9 mbar (3.3 mbar when the beacon is present) with a FWHM...
range from 0.4 to 9.4 mbar. These changes in altitude sensitivity depending on the atmospheric temperature are fully accounted for in the retrieval process. As the hydrocarbon contribution functions extend into the lower stratosphere to altitudes not covered by the CH₄ ν₄ emission (Fig. 2), we found it necessary to update our T(p) profiles very slightly while simultaneously scaling the hydrocarbon profiles to fit the 680–860 cm⁻¹ spectrum. The end product is a vertical T(p) profile (0.5 < p < 230 mbar) plus estimates of (i) the ethane and acetylene abundances near 2 mbar; and (ii) the phosphine gradient in the 100 < p < 550 mbar range.

3. Observations: the evolution of the beacon

Although the roiling convective activity in the troposphere had largely subsided by July 2011 (Sánchez-Lavega et al., 2012; Sayanagi et al., 2012), the stratospheric aftermath continued to the time of writing. The bright stratospheric airmasses dominated Saturn’s appearance at 7.9, 8.6, 12.3 and 13.0 μm (as shown in the VLT/VISIR imaging in Fig. 1), and the beacons could only be studied via their infrared emission which is mapped with latitude and longitude in Fig. 3. These figures consist of both VLT/VISIR 12.3-μm images (these were of a higher signal to noise than the 7.9-μm images) and reprojected Cassini/CIRS data averaged between 1290 and 1310 cm⁻¹ (the peak of the ν₄ emission band of CH₄), showing the westward motion of the beacons. The longitude of the beacons (in System III West) is recorded from Cassini, VLT and IRTF in Fig. 4. The maximum brightness temperature averaged over the 1290–1310 cm⁻¹ spectral range from the CIRS observations is shown in Fig. 5a, showing the variation in the intensity of the beacons. Uncertainties in the emission strength are standard deviations on the mean estimated within 5° of the beacon centre.

The longitudinal and latitudinal width of the warm airmasses is rather subjective, as it depends upon the selected ‘boundary’ of the beacons (we shall see in Section 4 that more accurate widths can be estimated from the maximal thermal windshear). In reality there is a continuous temperature gradient between the beacon core and the exterior, with no well-defined boundary to define a size. We elected to take the FWHM of the beacon compared to the quiescent background to demonstrate general trends in the morphology, and we plot the longitudinal width in Fig. 5b and the latitudinal extent of the beacon in Fig. 5c.

The evolutionary sequence can be divided into three phases: (Phase I) the formation and intensification of two distinct warm airmasses between January and April 2011; (Phase II) the merging and deceleration to form the large single beacon between April and June 2011; and (Phase III) the acceleration and slow cooling of the beacon since July 2011. Each phase is described below, and our results are summarised in Table 5.

3.1. Phase I: formation and intensification

The first Cassini thermal-IR observations of the storm latitudes occurred on January 2nd 2011, a month after the onset of the disturbance (Fig. 3). Previous observations in October 2010 showed longitudinal homogeneity, with no perturbations to stratospheric temperatures exceeding 2 K. Two warm airmasses were observed in January flanking a central cool region, with longitudinal brightness temperature contrasts of 13–15 K at 7.7 μm. Cassini/CIRS captured only the northern edges of the two airmasses at 50–55°N on January 2nd, but later observed the centres of the warm airmasses with a crude map on January 19th–20th.

The VLT/VISIR 8.6- and 12.3-μm images on January 19th (Fig. 1) confirmed the presence of a central cool stratosphere flanked at the eastern and western edges by warm airmasses. The western airmass, hereafter beacon one (B1), appeared more diffuse and extended, sitting above the bulbul white storm head observed in visible imaging. By comparison, the eastern airmass (beacon two, B2) was smaller and more confined. In these early stages, B2 appeared to be divided into two distinct warm regions in the VISIR imaging – one was over the ‘northern branch’ of the storm tail and was further to the east; the other existed over the ‘southern branch’ of the tail – but this split morphology was only observed once (January 19th 2011). I. During January 27th, the next VISIR observation in Figs. 1 and 3, the two airmasses comprising B2 had merged to a single warm airmass. By late January it was clear that the distance between B1 and B2 was increasing, suggesting different velocities for the two beacons.

The February 8th VISIR observation caught B2 close to the planetary limb, where limb brightening effects prevented us from extracting a reliable measure of the beacon strength (Fig. 1). However, the next CIRS observations on March 4th (Fig. 3) demonstrated substantial warming of the two airmasses since their first discovery in January. Fig. 5a shows that the maximum brightness temperatures of the two beacons increased in an approximately linear fashion to reach a peak in May 2011.

The March CIRS and VLT observations also suggest that the slower of the two beacons, B2, was at a more southerly latitude than the faster and more extended beacon, B1. The March 25th VLT/VISIR observations were designed to be obtained over 6 h so that both hemispheres of the planet could be viewed (Fig. 1 and a cylindrical map in Fig. 3). They also provided one of our clearest views of the beacon morphologies: B2 was compact and near-circular, centred at 27.8 ± 2.6°N and extending from 18 to 38°N (Fig. 5c). B1 had a more unusual shape, extending from southwest to northeast and centred near 33.2 ± 2.1°N. B1 had formed its own ‘tail’ of warm air, centred on 50°N but extending from 45 to 55°N. The most northerly stratospheric perturbations occurred north of B1 at around 65–70°N. The beacons appeared to produce no perturbations to the cold north polar region poleward of 70°N (Fig. 1), which is emerging from the darkness of winter shadow for the first time in 15 years. The images of B1 on March 25th (Fig. 1) show considerable structure in the extended emission around the central warm spot – at least three ‘satellite’ spots can be seen to the northeast, merging into the extended tail. The fate of these satellites is uncertain, as they were not imaged again before B1 and B2 merged, despite observations of equal spatial resolution.

It is also worth noting that the contrast between the beacon and the eastern tail is smaller in the 12.3-μm image than in the 8.6-μm image (Fig. 1). Calculating contribution functions for each of the VISIR filters under the high-temperature conditions of the beacon, we find that the 12.3-μm filter senses slightly deeper pressures (0.9–9.0 mbar, peaking at 3.3 mbar) than the 8.6-μm filter (1.0–5.7 mbar, peaking at 2.5 mbar), and that the 8.6-μm filter has an order of magnitude lower sensitivity to changes in stratospheric temperature than the 12.3-μm filter. Similar differences are observed in Fig. 2. Although these differences appear small, they may suggest a vertical structure to the beacon and tail, to which we shall return in Section 4.

3.1.1. Velocities pre-merger

The Cassini and ground-based observations continued throughout March and April as the stratospheric emission from the two discrete beacons intensified, and B1 completed its traverse of a full 360° longitude circle. By this point, enough data were available to quantitatively assess the beacon velocities throughout this period. Extrapolating the motions of the beacons backwards to the time of the storm onset in Fig. 4, we find that B1 would have been at 248°W on December 5th 2010, and B2 at 209°W. The initial tropospheric storm erupted at 242°W Sánchez-Lavega et al. (2011), close to the extrapolated location of B1. We speculate that
these two warm airmasses were generated as a direct result of the disturbance at 242°W on December 5th 2010, later known as the 'storm head'.

Fig. 4 shows that B1 then moved west at 2.7 ± 0.1 deg/day, equivalent to a velocity of −26.9 ± 1.1 m/s at 33.2°N (the centre of B1). Co-plotting the longitudes of B1 in the stratosphere and the white storm head in the troposphere (Sayanagi et al., 2012), we see a remarkable co-location of the two features between January and April 2011 (Fig. 4). B1’s velocity is similar to the velocity of the storm head (−27.9 ± 0.1 m/s from ground-based observations, Sánchez-Lavega et al., 2012), and consistent with the storm head drift rate of 2.8 deg/day reported from Cassini/ISS measurements (Sayanagi et al., 2012), suggesting a strong coupling between the tropospheric disturbance and the stratospheric feature. By compar-
ison, the westward velocity of the small tropospheric anticyclone that formed east of the storm head was 0.91 deg/day (Sánchez-Lavega et al., 2012), which does not seem to correspond to the beacon velocities in Fig. 4. Plotting the beacon velocities alongside Saturn’s tropospheric jet speeds (Sánchez-Lavega et al., 2000) in Fig. 5d, we observe that both the beacon and the storm head move faster than the retrograde jet at 39°N by around 7 m/s, indicating an increase in retrograde velocity from the typical tropospheric cloud decks (around 500 mbar) to the elevated altitudes of the white storm head (around 200 mbar).

The second beacon, B2, at the more southerly latitude of 27.8 ± 2.6°N, also moved to the west but much more slowly (0.6 ± 0.1 deg/day, Fig. 4), equivalent to a westward velocity of −6.2 ± 0.7 m/s at this latitude (Fig. 5d). This difference between the velocities of B1 and B2 caused the two beacons to drift together, intersecting and merging in late April 2011.

3.2. Phase II: merger

Fig. 4 shows that the two beacons occupied the same longitude on April 30th (day 120, ±3 days). Low spatial-resolution IRTF/MIRSI observations acquired in late April 2011 showed a single, large hot airmass (hereafter B0), with no distinct boundaries between the two independent beacons. The April 26th CIRS MIRTMAP (Fig. 3) shows an extended region of emission, occupying at least 80° longitude, with the warmest region at 280°W on that date. The remnant of B1 could still be seen as a secondary peak to the east of the hottest region, but this had been subsumed into B0 by May 5th, when it was clear that the interaction between the beacons had formed a single large hot airmass at 39.5 ± 4.1°N, further north than either B1 or B2 and near the centre of the westward jet. The width of the warm airmass, as specified by the FWHM (Fig. 5b), spanned some 90 ± 10° longitude immediately post-merger, and the temperatures reached their maximum. The maximum brightness temperature recorded at 1304 cm⁻¹ was 209.2 K on May 5th 2011 at a spectral resolution of 2.5 cm⁻¹, which is an increase of 70 K compared to the typical 140 K brightness temperatures of this latitude circle (physical temperatures will be discussed in Section 4).

From May 2011 onwards, the beacon was a discrete oval-shaped hot airmass (Fig. 3), wider in east–west extent (approximately 70° wide, equivalent to 55,000 km at 40°N, Fig. 5b) than north–south extent (covering approximately 30° in latitude, Fig. 5c). It continued to propagate westward, but more slowly than before, having become detached from the bulbous storm head beneath. Between April 30th and July 9th 2011 B0 detached from the storm head and moved west at 1.6 ± 0.2 deg/day (−14.6 ± 1.9 m/s at 39.5°N), a velocity intermediate between the two original beacons B1 and B2.

3.3. Phase III: mature beacon

The final ‘mature’ phase of the beacon B0 is defined to start in late June or early July 2011 (the exact date is ill-defined due to the scatter of points in Fig. 4). Visible observations showed that the white storm head had encountered the cold anticyclonic vortex in the troposphere between June 15th and 19th 2011 (Sánchez-Lavega et al., 2012). The bulbous white head had dissipated by July 12th, and active convection appeared to have ceased, leaving a dark tropospheric zone surrounded by disturbed cloud patterns to the north and south. The coupling between this convection and the beacon velocity is uncertain, but Fig. 4 indicates that B0 accelerated in late June or early July to a faster drift velocity of 2.70 ± 0.04 deg/day (−24.5 ± 0.4 m/s at 39.5°N). To calculate the...
### Table 5

Timeline of key events for the beacon evolution.

<table>
<thead>
<tr>
<th>Date</th>
<th>Event</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Phase I: Forma</strong></td>
<td><strong>tion and intensification</strong></td>
<td></td>
</tr>
<tr>
<td>05-December-2010</td>
<td>Tropospheric Storm Eruption at (37.7 ± 0.8)°N</td>
<td></td>
</tr>
<tr>
<td>02-January-2011</td>
<td>First detection of B1 and B2 by Cassini/CIRS</td>
<td></td>
</tr>
<tr>
<td>19-January-2011</td>
<td>First troposphere/stratosphere thermal maps of storm region by VLT/</td>
<td></td>
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<tr>
<td></td>
<td>VISIR</td>
<td></td>
</tr>
<tr>
<td>29-January-2011</td>
<td>Storm head encounters southern branch of the tail from the east</td>
<td></td>
</tr>
<tr>
<td>March 2011</td>
<td>Bright cloud material from the storm head encircled the planet</td>
<td></td>
</tr>
<tr>
<td>04-March-2011</td>
<td>First CIRS observations since January</td>
<td></td>
</tr>
<tr>
<td>25-March-2011</td>
<td>VLT/VISIR maps both B1 and B2 at high resolution</td>
<td></td>
</tr>
<tr>
<td>January–April 2011</td>
<td>B1 moves at 2.73 ± 0.1 deg/day; B2 moves at 0.6 ± 0.1 deg/day</td>
<td>Different drift velocities lead to merger</td>
</tr>
<tr>
<td>January–April 2011</td>
<td>Beacons intensify</td>
<td>Temperatures and C2H2 abundances rise</td>
</tr>
<tr>
<td>Late April 2011</td>
<td>B1 and B2 encounter each other near 300°W</td>
<td>Beacons merge</td>
</tr>
<tr>
<td><strong>Phase II: Merger</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>05-May-2011</td>
<td>CIRS Observations show single beacon B0</td>
<td>Peak temperatures 221.6 ± 1.4 K at 2 mbar recorded</td>
</tr>
<tr>
<td>May 2011</td>
<td>Merger accompanied by a downward shift of the temperature peak</td>
<td>Peak moved from 0.5 mbar to 2 mbar</td>
</tr>
<tr>
<td>April–June 2011</td>
<td>B0 detached from tropospheric storm head</td>
<td>Phase II velocity intermediate between B1 and B2 (1.6 ± 0.2 deg/day)</td>
</tr>
<tr>
<td>June 15th–19th 2011</td>
<td>Storm head encountered tropospheric anticyclone from the east.</td>
<td>Electrostatic discharges decreased, storm head dissipated</td>
</tr>
<tr>
<td><strong>Phase III: Mature phase</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>June-July 2011</td>
<td>B0 accelerates to phase III velocity 2.70 ± 0.04 deg/day (–24.5 ± 0.4 m/s at 39.5°N)</td>
<td>Remains constant for rest of observations</td>
</tr>
<tr>
<td>16-August-2011</td>
<td>First indications that B0 is longitudinally shrinking</td>
<td></td>
</tr>
<tr>
<td>Late September 2011</td>
<td>B0 completes first traverse around planet since merger</td>
<td></td>
</tr>
<tr>
<td>Mid January 2012</td>
<td>B0 completes second traverse around planet since merger</td>
<td></td>
</tr>
<tr>
<td>Early May 2012</td>
<td>B0 completes third traverse around planet since merger</td>
<td></td>
</tr>
<tr>
<td>July 2011–May 2012</td>
<td>B0 shrinks longitudinally (0.16 ± 0.01 deg/day) and cools (0.11 ± 0.01 K/day)</td>
<td>Remains at the same altitude (2 mbar) and C2H2 enhancement unchanged</td>
</tr>
</tbody>
</table>
System III west longitude, $\phi_{BO}$ (modulo 360), of the beacon on all dates beyond July 1st 2011, we used the formula:

$$\phi_{BO} = (2.70 \pm 0.04)t - (122.9 \pm 12.0)$$  \hspace{1cm} (1)$$

where $t$ is the reduced Julian date for January 1st 2011 (e.g., the Julian date minus 2466652.5). It remained at this velocity, some 4 m/s faster than the typical tropospheric retrograde jet (−20.4 m/s, Sánchez-Lavega et al., 2000) but slightly slower than the original storm head (−27.9 ± 0.1 m/s from ground-based observations, Sánchez-Lavega et al., 2011), for the remainder of 2011. The beacon velocity was also moderately faster than that of the original 'String of Pearls' (Momary et al., 2006), thought to be precursors of the eruption at this latitude (2.28 deg/day or 22.4 ± 0.2 m/s, Sayanagi et al., 2012).

B0 evolved during the second half of 2011, showing slow signs of dissipating. From August 16th onwards (DOY228), Fig. 5b shows that B0 began to show signs of longitudinal shrinkage and circularisation, to a FWHM of 30° longitude (23,400 km) by March 15th 2012 (the latitudinal extent in Fig. 5c remained approximately constant). Furthermore, the peak brightness temperature was cooling – a simple linear fit suggests a rate of 0.1 ± 0.05 K/day, or 36.5 ± 18.3 K/year (Earth years). At this rate, it will take approximately 620 days (1.7 Earth years) to return to quiescent temperatures after the beacon merger (January 2013). However, the decay of the beacon may be more rapid if the feature continues to shrink in size.

In summary, the brightness temperature maps have been used to track the location of these unique stratospheric features from their original discovery, through their intensification and merger and into a mature phase of cooling and shrinkage (Table 5). In the following sections, we will use Cassini/CIRS spectroscopy to investigate the vertical dimension of the stratospheric beacons, in terms of their temperature structure and composition.

4. Results: beacon temperatures and composition

Each of the pixels in the Cassini/CIRS maps (Fig. 3) represents a complete spectrum from 600 to 1400 cm$^{-1}$, which can be used to derive the thermal structure and stratospheric composition to add a vertical dimension to the horizontal maps. Examples of the beacon spectra at the highest CIRS spectral resolution of 0.5 cm$^{-1}$ compared to the quiescent background are shown in Fig. 6a for July 7th 2011, during the mature phase of B0. The brightness temperatures of the CH$_4$ $v_2$ band are 30–40 K greater in the beacon, and the emission lines appear broadened within B0. These high temperatures also increase the emission from a host of minor hydrocarbon species in Fig. 6b, notably ethane ($C_2H_6$) at 825 cm$^{-1}$ and 1390 cm$^{-1}$ and acetylene ($C_2H_2$) at 730 cm$^{-1}$, which will be the subject of modelling in Section 4.4. But we also observe enhanced emission from species such as propane ($C_3H_8$), methyl-acetylene ($C_3H_4$), diacetylene ($C_4H_2$), ethylene ($C_2H_4$, studied by Hesman et al. (2012)) and CO$_2$, which will all be the subject of future studies.

4.1. Vertical and longitudinal temperatures

For each CIRS observation in Table 1 that passed through the centre of the hot airmasses, we coadded spectra between 30° and 45°N on a longitude grid with 10°-wide bins and stepped every 5°, to retrieve longitudinal cross sections of temperature and composition (Fig. 7). In addition, we coadded data within ±10° longitude and ±5° latitude of the centres of the beacons to generate a selection of individual $T(p)$ profiles shown in Fig. 8. As COMPSIT (0.5 cm$^{-1}$ spectral resolution) data featured fewer spectra in each coadd than the MIRMAP (2.5 cm$^{-1}$ resolution) data, we applied a Hamming convolution function to smooth the COMPSIT data to the MIRMAP spectral resolution.

4.1.1. Temperatures during formation and merger

Initial observations: Pre-storm longitudinal $T(p)$ cross sections acquired on November 4th 2009 (a FIRMAP), 12 months before the storm outbreak, revealed 0.5-mbar contrasts no larger than 4 ± 2 K, and no contrasts in the troposphere larger than the retrieval uncertainties (mean 110-mbar temperature of 83.5 K). The first three plots of Fig. 7 were originally published by Fletcher et al. (2011), which compared MIRMAP 'shift-and-stare' observations on October 22nd 2010 and January 20th 2011 (pre- and post-storm), showing a cool stratospheric airmass (137 ± 3 K at 0.5 mbar, similar to the quiescent temperatures of 140 ± 2 K) centred on 300 W above the cold anticyclonic vortex in the troposphere, and flanked by warm regions to the east and west. These warm regions had been first glimpsed in a more northerly (45°–55°N) COMPSIT on January 2nd 2011 (Fig. 7), with peak temperatures in the 0.4–0.6 mbar range and contrasts between B1 (151 ± 2 K), B2 (148 ± 2 K) and the unperturbed stratosphere (140 ± 2 K) were 8–11 K. Contrasts were smaller (5–8 K) at 5 mbar, but no evidence for these hot airmasses was seen at the tropopause (100 mbar). The January 20th observations in Fig. 7, 45 days after the storm outbreak, revealed 0.5-mbar contrasts between B2 and the disturbance centre of 20 ± 3 K, and between B1 and the disturbance centre of 12 ± 3 K. At this time, these were the largest zonal
Fig. 7. Longitude-pressure cross sections of upper tropospheric and stratospheric temperatures throughout 2011 and 2012. The temperature scale is the same in all panels, partially masking the two beacons during January 2011. Spectra were coadded between 30° and 50°N. Blank sections indicate an absence of data at these longitudes. The March 15th observation failed to capture the centre of B2, which is why the temperature perturbation appears smaller. The drop in altitude during the merger can be seen by comparing 2011–March–04 with 2011–October–22. Post-merger, the altitude remains approximately constant but the longitudinal extent and peak temperatures cool with time. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
thermal anomalies ever detected in Saturn’s stratosphere (Fletcher et al., 2011).

Pre-merger phase I: Our best example of B1 and B2 in their pre-merger phase came on March 4th 2011 with a 0.5 cm⁻¹ COMPSIT observation that captured the peak temperatures of both beacons. Fig. 7 shows that B1 and B2 remained centred at the 0.5-mbar pressure level, with peak temperatures of 190 ± 1 K and 188 ± 1 K respectively, compared to a quiescent stratospheric temperature of 140 ± 3 K. B2 appeared more compact and symmetric about its central longitude, whereas B1 had an extended shape and resided directly above the storm head in Fig. 3. At the 0.1-mbar level, where the information content of nadir CIRS spectra is greatly diminished, there is evidence that the stratosphere is cooler immediately above the beacons (seen clearly in Fig. 8a) and surrounded by an extended region of warm emission, broader in longitude at 1 mb pressures than at mbar pressures. This hints at the possibility that the beacons may have spatial structure at lower pressures than CIRS can reliably measure. At higher pressures beneath the beacons, the warm thermal anomaly extended to the tropopause at 70–80 mbar (the temperature minimum is warmer beneath the beacons than elsewhere, Fig. 8a). The coldest stratospheric region was observed immediately west of B2 (centred at 310°W on March 4th, Fig. 7), and appeared significantly cooler than the ambient atmosphere (134 ± 2 K at 1 mbar compared to ‘unperturbed’ temperatures of 140 ± 3 K). This cold stratosphere is related to a cold tropopause region in Fig. 7, but this does not correspond to any notable storm features on this date (neither the storm head nor the tropospheric oval).

Merger phase II: CIRS obtained 2.5 cm⁻¹ spectra of B1 and B2 as they merged to form B0 on April 26th and May 5th, at which point the peak temperatures exhibited a downward shift to the 2 ± 1 mbar level, and produced an increasingly large perturbation to the upper troposphere (Fig. 7). The structure of the longitudinal cross section became complex, with B0 appearing narrower at high pressures, and broader at lower pressures. Immediately after merger the B0 complex appeared to be asymmetric, with a more ‘north–south’ boundary to the west, and a sloping ‘southwest–to–northeast’ gradient to the east, possibly associated with the warm tail of emission observed in Fig. 3. The 110-mbar temperatures were 7 ± 3 K warmer than regions away from the beacon centre. The 0.5-mbar temperature cross-section showed a uniform temperature of 180–190 K over 80° longitude, whereas the peak temperatures of the core (295°W on May 5th 2011) were evident deeper at 2 mbar. The maximum temperature retrieved was 221.6 ± 1.4 K at 2 mbar on May 5th and 218.0 ± 1.5 K at 1.7 mbar on April 26th, approximately 80 K warmer than the quiescent background atmosphere (mean temperatures of 140 ± 3 K at 2 mbar).

4.1.2. Temperatures during decay phase III

From May 5th onwards, the peak temperature of B0 remained at 2 ± 1 mbar (i.e., deeper than the 0.5-mbar peaks of B1 and B2) in Fig. 8a. The right hand side of Fig. 7 shows that the beacon became more symmetric from July 2011 to March 2012, and began to show signs of cooling. Within the beacon itself, and taking 1 bar as the 0 km reference level, the 2-mbar level is at 225 km, the 0.5-mbar level is at 310 km, representing an altitude shift for the peak.
temperatures of approximately 85 km, equivalent to \( \ln(2/0.5) \approx 1.4 \) scale heights. Representative spectra in Fig. 9 demonstrate the change in the appearance of the \( \nu_4 \) \( \text{CH}_4 \) emission band from pre- to post-merger due to this altitude shift. The pre-merger spectra are characterised by a strong central Q-branch and weaker P and R branches, with narrow emission features. Post-merger, the emission features increased in width (suggesting enhanced pressure broadening deeper in the atmosphere), and the significance of the Q branch diminished, consistent with the downward motion of the beacons’ peak temperatures in Fig. 8a.

Surprisingly, the peak 2-mbar temperatures were observed to vary very slowly over the timespan of these observations, from 196.5 ± 1.1 K on July 8th, to 193.2 ± 1.1 K on September 10th 2011, 194.3 ± 1.2 K on December 4th (consistent with September observations to within the error); 191.8 ± 1.0 K on January 20th 2012 and 186.1 ± 1.1 K on March 15th 2012. In total, the 2-mbar temperatures decreased by 35.5 ± 2.5 K between May 4th 2011 and March 15th 2012 (316 days), or a rate of change of 0.11 ± 0.01 K/day. At this linear rate, it would take 2 years to cool to the quiescent temperature of 140 K. A similar estimate of 1.7 years was obtained from the raw brightness temperatures alone. For comparison, the radiative time constant for Saturn’s nominal atmosphere at 2 mbar is approximately 10 years (Conrath et al., 1990).

Despite this rather slow decline of the peak temperatures, the morphology of the beacon in Fig. 7 was changing, with the longitudinal extent shrinking. If we arbitrarily select the 160 K and 170 K contours, we can plot the width of B0 for each longitudinal cross-section (Fig. 10). Only dates with full coverage of the beacon are included here, and a conservative error of ±5° has been applied to account for the confusion generated by the extended northeastern ‘tail’ of the beacon. The two contours show a similar
downward trend since April 26th 2011, with the 160-K contour being 88 ± 5° longitude wide (=69,000 km) during and immediately after the formation of B0, but having decreased by a factor of two to 38 ± 5° longitude (≈30,000 km) by March 15th 2012. Cassini/CIRS measurements on December 4th 2011 appear unusual compared to the others in Fig. 7 because the eastern tail of emission (captured by the 160-K contour) appeared more extended on this date (see Fig. 3). A linear fit through the 160-K contour widths suggest shrinkage by 0.16 ± 0.01 deg/day, such that it would reach zero 660 ± 100 days after January 1st 2011, or 1.8 ± 0.3 years (i.e., in the second half of 2012). This is consistent within the error bars with the shrinkage of the beacon FWHM of its CH₄ emission in Fig. 3. The beacon therefore appears to be shrinking in spatial extent at a faster rate than the peak temperatures are cooling.

4.1.3. Atmospheric stability

Given the presence of the large hot airmass in Saturn’s stratosphere, it is appropriate to consider how this modifies the atmospheric convective stability in Fig. 8. Fig. 8b shows the variation in the atmospheric lapse rate dT/dz, confirming that the lapse rate becomes negative in the stratosphere above the beacon centre (0.5-mbar pre-merger and 2-mbar post-merger), but that the lapse rate never exceeds the dry adiabatic lapse rate, (0.5-mbar pre-merger and 2-mbar post-merger), but that the lapse rate becomes negative in the stratosphere above the beacon centre (0.5-mbar pre-merger and 2-mbar post-merger). The peak temperatures of the tail (175 ± 2 K at 0.5 mbar) were located between 48 and 50°N at a pressure of 0.3–0.8 mbar, but the broken nature of the band makes its latitudinal extent difficult to quantify. The warm band appears to lie between the prograde tropospheric jet at 47.2°N and the retrograde jet at 55.1°N, although this correspondence is not precise, and the change in the north-south gradient of temperatures associated with this stratospheric band will alter the vertical windshear on the zonal jets. Longitudinal cross-sections at 1 mbar through Fig. 11 showed east–west contrasts of 28 ± 2 K through the warm band (55°N), compared to 50 ± 3 K for B0 (40°N), and 15 ± 2 K for the most southerly portion of the disturbed stratosphere (15°N).

Fig. 12 shows that B0 extends from 20 to 60°N, with broader temperature contours (i.e., larger north–south extent) at higher altitudes. The location of the temperature maximum is deepest (2 mbar) within the core of the beacon near 40°N, but rises upwards to 0.3–0.8 mbar pressure levels as we move northwards into the tail region. However, these high altitudes are beyond the capabilities of nadir remote sensing to provide reliable T(p) retrievals. In Section 3, we noted that VISIR imaging of the beacon and the warm northern band showed different contrasts depending on what filter (8.6, 12.3 or 13.0 μm) was being used. In Fig. 12 we can now determine why this should be the case – B0’s peak temperatures are at the 2-mbar level, whereas the peak temperature of the tail is at higher altitudes. Tropopause temperatures (100 mbar) in Fig. 11 show the expected structure of warm belts and cool zones, and are only affected beneath B0 (with longitudinal thermal contrasts of 4 ± 2 K), and not beneath the new stratospheric band.

Finally, Fig. 12 also shows that the equatorial stratosphere is cooler than 10°N in August 2011, consistent with the continued downward propagation of the thermal structures associated with Saturn’s Semianurnal Oscillation (SSAO, Orton et al., 2008; Fouchet et al., 2008; Fletcher et al., 2010; Guerlet et al., 2011; Friedson and Moses, 2012). Furthermore, the stratospheric temperatures show a minimum near 77°N (the latitude of Saturn’s hexagonal wave in the troposphere), indicating the subtle warming of the north polar stratosphere during spring, initially observed in wintertime conditions in 2007 (Fletcher et al., 2008), but the north polar stratosphere remains cooler than any other latitude.

4.4. Windspeeds and beacon circulation

The primitive equations of atmospheric motion (e.g., Holton, 2004) allow us to relate the horizontal temperature variations to the zonal, meridional and vertical velocity fields. Vertical wind shears are calculated by assuming geostrophic balance between
Fig. 11. Horizontal variations of beacon properties during the mature phase III, retrieved from 15-cm⁻¹ spectral resolution mapping observations on August 21st 2011. Panels a–c show retrieved temperatures at the 0.5-, 2- and 100-mbar levels. Panels d and e show the calculated stratospheric winds from an integration of the thermal windshear equations. Panel f computes the magnitude of the winds ($\sqrt{u^2 + v^2}$) showing the peripheral collar of the vortex. Panels g and h show the spatial distribution of acetylene and ethane at 2 mbar. The key for each panel is shown to its right.
Fig. 12. The latitudinal variation of temperatures through the beacon, averaged over 10° of longitude at three different locations on August 21st 2011 (FIRMAP observation). A latitude-pressure temperature cross-section from a 2009 observation is shown at the top of the figure for comparison. The three rows beneath represent longitudes of 310, 140 and 40°W, through the beacon tail, the centre of B0, and the cold region to its east in Fig. 3, respectively. On the left we show the zonal temperatures, and on the right we show the temperature difference compared to the 2009 observation. The same temperature scale is used for all similar plots. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)
the meridional pressure gradients and Coriolis forces. In log-pressure coordinates, the thermal windshear equations for the zonal \((u)\) and meridional \((v)\) directions are:

\[
f \frac{\partial u}{\partial \ln(p)} = \frac{R}{a} \frac{\partial T}{\partial y} = -R \frac{\partial T}{\partial \lambda}
\]

\[
f \frac{\partial v}{\partial \ln(p)} = -\frac{R}{a \cos \psi} \frac{\partial T}{\partial \phi} = -R \frac{\partial T}{\partial \phi}
\]

where \(T\) is the temperature in Kelvin; \(p\) is the pressure in bar; \(f\) is the Coriolis parameter \(f = 2\Omega \sin(\psi)\) where \(\Omega\) is the planetary angular velocity, \(\psi\) is the latitude; \(a\) is the mean planetary radius; \(\phi\) is the longitude and \(R\) is the molar gas constant divided by the mean molar weight of Saturn’s atmosphere. If we take 400 m/s as the peripheral velocity \(U\) (Fig. 11), and 50° longitude (approximately 40,000 km) as a mean length scale \(L\), then the Rossby number \((R_o = U/\ell L)\) is \(\approx 0.05\), small enough that geostrophy is maintained within the beacon.

The horizontal variation of the zonal and meridional stratospheric winds is presented in Fig. 11d–f at 2 mbar for the August FIRMAP. If we assume the tropospheric zonal wind field of Sánchez-Lavega et al. (2000) is referenced to the 500-mbar level, then these vertical windshears serve to accelerate zonal velocities at the 2-mbar level that exceed \(300\) m/s at both the northern (prograde, largest \(dT/\partial y\) between 48 and 50° N) and southern (retrograde, largest \(dT/\partial y\) between 25 and 25° N) boundaries of B0 (Fig. 11). In the meridional direction, we assume a level of no motion at the tropopause, and integrating the windshear equations produces 2-mbar northward velocities of \(v > 200\) m/s at the western boundary (175°W) and southward velocities of \(v < -200\) m/s at the eastern boundary (110°W). Taken together, the peripheral velocities in Fig. 11 suggest a clockwise circulation around B0, with the largest magnitude \((\approx 400\) m/s) in a collar surrounding the vortex core, making this an anticyclone in the northern hemisphere. In this calculation, the meridional velocities appear to be smaller than the zonal velocities, something also observed in the circulation of Jupiter’s Great Red Spot (Choi et al., 2007), where maximum tangential velocities of \(170\) m/s were measured in the collar. To confirm the rotation of the vortex, the relative vorticity, a measure of the local change of the fluid (the curl of the velocity field) is calculated as \(\zeta = dv/\partial x - du/\partial y\). Considering Fig. 11, we can see that \(dv/\partial x\) (west to east) is strongly negative and \(du/\partial y\) (south to north) is positive, resulting in an overall negative \((i.e.,\) anticyclonic) relative vorticity within the beacon.

The meridional velocities before the storm, pre-merger and post-merger are compared in Fig. 13. When we calculate the meridional velocities for the March 4th 2011 COMPSIT observation, which captured both B1 and B2, we find 1-mbar meridional velocities for B2 of 140 m/s northward at 280°W and 80 m/s southward at 250°W, and for B1 we find 80 m/s northward at 130°W and 70 m/s southward at 110°W (Fig. 14). The location of the maximum \(dT/\partial x\) varies with height, as suggested by the longitudinal temperature cross-section in Fig. 7, but this result suggests that the two original beacons, B1 and B2, were anticyclonic vortices some 20–30° wide with peripheral clockwise velocities of 70–140 m/s (the broad range represents the large uncertainties associated with extrapolating the thermal winds to stratospheric altitudes, assuming a level of zero meridional motion at 100 mbar). Fig. 14 shows that the peripheral velocities around the combined B0 were greater at the 1-mbar level than the individual beacons in March 2011, consistent with the downward shift of the peak temperatures from 0.5-mbar to 2 mbar after the merger.

Finally, we investigate the zonal wind shear introduced by the presence of the new warm stratospheric band at 48–50°N. If we compute the zonal winds using a north–south cross-section at 310°W (i.e., through the beacon tail but not the main vortex, Fig. 11), we find prograde accelerations (80–120 m/s) at 52°N and 60°N and a retrograde acceleration (120–160 m/s at 47°N) at 1-mbar, bounding the tail of emission. Thus the new warm band at the northern edge of B0 could be associated with zonal jets in the stratosphere, prograde poleward of the new warm band and retrograde on the equatorward side, provided some process is able to transport momentum (either by advection or eddy fluxes) into these stratospheric jets.

### 4.4. Compositional variability

The peripheral jets surrounding the stratospheric anticyclone raise the possibility that the air mass within B0 is isolated from the ambient stratosphere, permitting unique environmental conditions. As described in Section 2.4, we varied the molecular abundances of ethane (C\(_2\)H\(_6\)) and acetylene (C\(_2\)H\(_2\)) as we retrieved the vertical \(T(p)\). This simultaneous fitting of temperatures and hydrocarbons gave slightly closer fits to the ethane and acetylene bands than a two-stage process, as the \(T(p)\) derived from the CH\(_4\) band required minor adjustment within the retrieval uncertainties to reproduce the hydrocarbon emission. As their photochemical lifetimes are long relative to the lifetime of the storm, spatial variations in hydrocarbon mixing ratios can be used to infer the timescales for vertical transport by mixing in the stratosphere (e.g., Moses and Greathouse, 2005). In this study we make no attempt to derive the details of the vertical profiles of these species, but we retrieve a simple scaling factor for an \(a\) priori profile based on mean hydrocarbon distributions from Cassini/CIRS limb analyses (Guerlet et al., 2009). In this way we can show trends in the molecular abundances, but we avoid issues related to non-overlapping contribution functions from CH\(_4\), C\(_2\)H\(_2\) and C\(_2\)H\(_6\) emission features (see Section 2.4).

#### 4.4.1. Acetylene

In Fig. 15, we use two techniques to show the variation of C\(_2\)H\(_2\) with longitude and time – a cascade plot on the upper panel, with each successive date offset by 2 ppm, and a comparison plot in the lower panel (for post-merger only), with the beacon shifted to 180°W for all dates. On January 2nd 2011 we detect a contrast of 0.1 ppm between the storm core and the adjacent beacons at 310° and 230°W, at the level of the uncertainties on the retrieved abundance. This had increased by January 20th 2011, where B2 (260°W) was enhanced in acetylene compared to adjacent longitudes by around a factor of two (the cold tropospheric anticyclone was near 315°W on this date).

Variability in C\(_2\)H\(_2\) is better shown in Fig. 16a, where we used coadded spectra of the individual beacons on each date to determine the abundance of C\(_2\)H\(_2\) in their centres. Pre-merger, we found that B1 and B2 had peak abundances of 0.8 ppm at 1 mbar (March 4th 2011), and continued to grow (Fig. 16a), reaching a peak abundance of 1.6 ppm at 1 mbar on May 5th, shortly after the merger. Post-merger, the trends in Figs. 15 and 16a are more complex, with no obvious decline in the beacon abundance, remaining enhanced by a factor of three over quiescent conditions. Fig. 11g maps the C\(_2\)H\(_2\) distribution in August 2011, showing it to be uniformly enhanced within the boundary defined by the peripheral high-velocity collar. During the mature phase, the abundance varies around a mean of 1.3 ± 0.2 ppm (a factor of three greater than the typical mean abundance of 0.4 ± 0.05 ppm at this latitude), but it is likely that this variability is due to uncertainties arising from (i) inconsistent calibration between temporally separated CIRS observations and (ii) slight differences in the latitudes and longitudes contributing to the coadded spectra.
4.4.2. Ethane

Figs. 17 and 16b use the same techniques for showing the evolution of ethane. In general, the longitudinal distribution of ethane is more erratic, with east–west variability even in the absence of the storm perturbations (e.g., January 2011). By January 20th 2011, there are suggestions that ethane was enhanced within the beacons, but the contrasts between the warm vortex and the cooler surroundings is much smaller than for acetylene. In May 2011, immediately post-merger, there was evidence that ethane was increased to the east of B0. However, when we compare all observations of ethane after this date in the bottom panel of Fig. 17, there is little consistent evidence for a trend in the $\text{C}_2\text{H}_6$ abundance. This is confirmed by the spatial distribution of ethane in August 2011 (Fig. 11h), which shows the lack of ethane enhancements. Finally, if we retrieve $\text{C}_2\text{H}_6$ from the coadded spectra on each date, the results hint that ethane may be enhanced ($7.5 \pm 1.0$ ppm) in the beacon compared to the quiescent background ($6.5 \pm 0.5$ ppm), but the retrieval uncertainties are large.
4.4.3. Acetylene versus ethane

This poses the question of why acetylene shows such a strong enhancement within the beacon, whereas the effects for ethane are more subtle and within our uncertainties (e.g., Fig. 11). This difference in the behaviours of the hydrocarbons betrays the simplicity of our approach (scaling of a priori distributions), but two additional factors are likely to play a role: (i) the two species have different vertical gradients, and (ii) the line cores of C₂H₂, unlike C₂H₆, have sensitivity to \( \mu \)bar pressures (e.g., Fig. 2) where we are unable to independently constrain the temperatures.

Fig. 6 of Moses and Greathouse (2005) shows that both hydrocarbons have source regions in the upper stratosphere at \( \mu \)bar pressures, and that the vertical gradient of the hydrocarbon distributions is seasonally variable. However, at the altitudes of interest (0.1–10.0 mbar) there is little change with season, and ethane has a larger gradient than acetylene. Where ethane’s gradient is more linear with altitude (26 ppb/km), acetylene’s gradient varies largely over the range of interest, from 1 ppb/km near 5 mbar to 12 ppb/km near 0.1 mbar, making it harder to assess simple vertical translations of the profile. The vertical profiles of these hydrocarbons are likely being dynamically perturbed away from their reference profiles by the subsidence within the beacon, adjusting the vertical slope of the gas to make it more uniform with altitude and increasing the column abundance. Furthermore, the higher temperatures of the beacon may alter the typical photochemical reaction rates, although Fig. 13 of Moses and Greathouse...
(2005) demonstrates that the gradients of the hydrocarbon profiles have a limited sensitivity to the stratospheric temperatures (they are more sensitive to the seasonal variation of solar flux, for example). It is also interesting to note that C₃H₂ is an important product of ethylene photolysis (e.g., Moses et al., 2000), another hydrocarbon observed to be greatly enhanced within the beacon (Hesman et al., 2012).

To address the second possibility, we attempted to fit the C₂H₂ emission by varying T(p) alone, keeping the abundance at the ambient values. The double-peaked nature of the contribution function (Fig. 2) meant that the line cores could only be fitted by a second high-temperature beacon at μbar pressures, leaving an ambiguity between temperature and C₂H₂ abundance that cannot be resolved by CIRS (the CH₄ band used for temperature sounding is only sensitive to μbar pressures). The CH₄ band fits cannot fully rule out heating at μbar pressures and above as an alternative to the enhancement of C₂H₂. A resolution of these questions must await sounding of the vertical profiles via higher-spectral resolution datasets (e.g., those acquired by Hesman et al. (2012) and Greathouse et al. (2011)).

4.5. Absence from reflected sunlight observations

One puzzle concerning the hot stratospheric temperatures is the apparent absence of any effects in reflected sunlight, implying that the anticyclonic vortex is (i) having no effect on the tropospheric cloud decks; and (ii) no effects on the haze materials in the upper troposphere and lower stratosphere. To investigate this further, a full Cassini/VIMS 0.8–5.1 μm cube covering the beacon longitude on July 12th 2011 (63–16°W at a latitude of 15°N, with wider longitude coverage at more northerly latitudes) was provided by K. Baines and T. Momary (private communication). The cube was acquired at 21:54 UT on July 12th 2011 (number 1689201784). We found no structures in the clouds and hazes corresponding to the beacon, focussing on the strongly-absorbing bands to sense the lower stratospheric hazes. The most prominent feature was the low-reflectivity band between 25° and 45°N left in the wake of the storm. Indeed, there was nothing notable about the beacon longitude in reflected sunlight in the VIMS cubes.

We can also compare to Cassini/ISS data obtained on July 12th 2011 (Fig. 9 of Sayanagi et al. (2012)), which also covers a complete longitude circle. Their false-colour maps use a combination of CB2–MT2–MT3 filters, with methane filters probing upper-level hazes. Although the storm latitude band is replete with small-scale vortices and eddies, there is nothing notable about the beacon longitude.

Finally, we alerted the amateur community to the beacon longitudes throughout 2011 and 2012, to assess whether small-scale clouds and spots were more prevalent at the beacon longitudes than elsewhere. Submissions were received from T. Barry, A. Wesley, C. Go, P. Haese and M. Delacroix, and we continued to survey the Planetary Virtual Observatory Database (Hueso et al., 2010) for any signs of unusual activity. However, at the time of writing, we conclude that the hot stratospheric temperatures are having no effect on the clouds and hazes at higher pressures, and cannot be viewed in reflected sunlight. The most prominent feature in the visible, however, was the new white axisymmetric band between 45° and 50°N that had formed in the aftermath of the storm. The southern edge of this band was undulating with waves, corresponding to the continued churning of the storm band. In March 2012, this bright reflective band sits north of the low-reflectivity storm latitudes observed by VIMS (25–45°N), but slightly south of the warm tail of beacon emission identified previously between 50 and 60°N.

5. Discussion

The thermal-IR data have captured the formation, strengthening, merger and decay phase of a large warm anticyclonic vortex in Saturn’s northern stratosphere that is invisible in reflected sunlight (see summary in Table 5). Taking the vortex width to be defined by the maximal east–west temperature gradients, dT/dx, we find that the beacon was some 50,000 km wide in August 2011 (Fig. 11), a factor of two larger than Jupiter’s Great Red Spot. In this section we discuss theories for the origins, nature and motion of this new stratospheric vortex.

5.1. Beacon origins

The tropospheric storm started in the vicinity of Saturn’s first retrograde jet north of the equator at 39.2°N. These westward mid-latitude jets have been previously identified as sites of potential barotropic instability (Read et al., 2009; Achterberg and Flasar, 1996), leading to interesting dynamical phenomena and eddy activity derived from the mean zonal flow. For example, this is also the latitude of Saturn’s string of pearls, a chain of small cloud-free vortices first identified in 5-μm emission (Momary et al., 2006). Indeed, Sayanagi et al. (2012) suggest that the eruption occurred within the pearl features themselves. In the southern hemisphere, the first retrograde jet at 40.7°S is associated with the string of small-scale convective clouds and vortices known as storm alley. However, never before have these tropospheric perturbations produced detectable perturbations to Saturn’s stratospheric emission.

Reflected sunlight observations of the tropospheric cloud deck are consistent with the disturbance being initiated as a convective storm (or multiple convective plumes) driven by the condensation of water (Sánchez-Lavega et al., 2011; Fischer et al., 2011). The
release of latent heat due to water condensation has long been sus-
ppected as the source of energy for upward transport within giant
planet atmospheres (e.g., Gierasch et al., 2000). Although some de-
gree of convective overshooting to the 60-mbar level, 15–20 km
above the tropopause, was expected from moist convection models
(Sánchez-Lavega et al., 1999; Hueso and Sánchez-Lavega, 2004),
such large perturbations at the 1-mbar level cannot be attributed
to direct convection.

Instead, the initial formation of the warm airmasses is likely to
be a stratospheric response to the mechanical forcing from below,
with the strong convective plumes impinging on the tropopause
acting as a topographic forcing. On Earth, convective activity in

Fig. 17. Variation of ethane with longitude from January 2011 to March 2012, showing much smaller perturbations than that of acetylene. The upper panel shows a selection of longitudinal retrievals, offset by 6 ppm for each date. The lower panel shows all the post-merger observations on top of one another (i.e., shifting the beacon to 180° for all dates). With the exception of the May 2011 enhancement in ethane to the east of B0 (immediately after the merger), we observe little variation in the mole fraction outside of the retrieval error. All available spectra were coadded between 30° and 45°N for each date.

thunderstorms are known to be a source of gravity wave activity,
so waves initiated in Saturn’s troposphere may have been able to
extend vertically to perturb the quiescent stratosphere above the
storm head. The challenge is in identifying the nature of Saturn’s
storm waves and determining where and how such waves depos-
it their energy to produce the heating detected at or above the
0.5-mbar level (the upper limit to CIRS nadir sensitivity).

The merger of the two beacons to form the larger vortex is an
example of an inverse energy cascade from smaller to larger scales,
as one would expect from a weather system governed by two-
dimensional turbulence where vortices generated in shear regions
are expected to merge and grow. The maximum extent of vortices
typically depends on the width of the belt or zone in which they are embedded, but the abnormally large size of the stratospheric anticyclone, cutting across jet streams at 32°, 39°, 47° and 55°N, hints at a different regime of zonal motion in the high stratosphere. The reasons for the downward shift in the beacon post-merger, by 1.4 scale heights, and the associated enhancement in temperatures and acetylene abundances, are not understood.

5.1.1. Numerical simulation

Given previous successes in reproducing the storm morphology at Saturn’s cloud tops with non-linear simulations (e.g., Sánchez-Lavega et al., 2011), and in particular the tropospheric cold anticyclonic vortex observed by Fletcher et al. (2011), we attempted to simulate beacon formation using the EPIC atmospheric model (explicit planetary isentropic-coordinate, Dowling et al., 1998). EPIC is a primitive-equation model, and fully includes production of Rossby waves, Kelvin waves and gravity waves by the storm head, although modelling their properties correctly remains a challenge. The reference T(p) from Section 2.4 was divided into 10 vertical layers (isentropic surfaces) between approximately 0.01 mbar and 10 bar. A steady vertical Gaussian heat source exists for p > 200 mbar to mimic the powerful convective activity of the storm head (lowest three layers of the model), whose vertical extent, horizontal width and intensity can be varied (nominal intensity of 1000 W/m²). This heat source is an intense source of gravity waves, which propagate outwards in three dimensions, horizontally and vertically through the stably-stratified upper atmosphere. However, the waves were reflected at the boundaries of the domain (despite the use of ‘sponge layers’ in the top two layers of the model) and result in temperature maps by day 30 that were dominated by noise and bore no resemblance to the Cassini measurements. EPIC cannot provide information on where these waves ultimately deposit their energy, a limitation of the horizontal and vertical domain ranges selected to allow for reasonable computational efficiency.

The natural output of EPIC are maps of potential vorticity (PV), which can be considered as a passive tracer of the dynamics. Our simulations confirmed that tropospheric activity can produce perturbations to PV distributions at stratospheric altitudes, but the morphology of the PV field at 1 mbar is extremely sensitive to the chosen vertical windshear, du/dz, which is poorly constrained by measurements (see Appendix A). Simulations often produced large-scale stratospheric airmasses with anticyclonic vorticity (reminiscent of B1) in the vicinity of the upwelling and divergence over the tropospheric storm head. However, definitive conclusions are not possible due to the sensitivity to the chosen u(z); the limited domain size and gravity wave internal reflections; and the coarse nature of the vertical grid. Improvements of these simulations will be the subject of future research.

5.1.2. Relationship to planetary waves

One possible interpretation of the beacon origins is that they were formed as topographically-forced Rossby waves due to the powerful tropospheric convection. Charney and Drazin (1961) demonstrated that the vertical propagation of planetary wave disturbances from the troposphere (caused by mechanical forcing from instabilities) depends on the effective index of refraction of the waves, and hence on the variation of the mean zonal wind with altitude. On Earth, strong Rossby-wave disturbances in the stratosphere are frequently due to upward propagation of waves generated by large-scale weather disturbances in the troposphere (e.g., Andrews et al., 1987). Non-linear vertical eddy transports of heat and momentum associated with these waves could then modify the basic zonal flow and produce the stratospheric perturbations we see. When the flow is prograde, the atmosphere acts as a short-wave filter and only the longest wavelengths of Rossby waves can propagate vertically, meaning that the stratospheric disturbances are large in scale. Achterberg and Flasar (1996) previously identified 37°–41°N as a critical region where the potential vorticity gradient vanishes and instabilities could excite Rossby waves.

Rossby waves owe their existence to the Coriolis force with latitude (known as the β effect, where β = d(2Ω sin ψ)/dy = 2Ω cos ψ/Rf is the Coriolis parameter 2Ω sin ψ, ψ is the latitude; R is the planetary radius and f is the planetary rotation velocity), which acts as a restoring force for parcels displaced meridionally. In a three-dimensional atmosphere, the zonal phase speed c relative to the mean zonal wind u is given by (Sánchez-Lavega, 2011):

\[ u - c = \frac{\beta}{k^2 + f^2 + (f_0^2/N_b^2)(m^2 + 1/4H^2)} \]

(4)

Here k, l and m are the zonal, meridional and vertical wavenumbers, respectively; H is the atmospheric scale height (50.2 km for the latitude and altitude of the beacon); and N_b is the Brunt Väisälä frequency (considered to be constant at 7 × 10^{-3} s^{-1} at the altitudes of the beacons). The Coriolis parameter is f_0(40°) = 2.1 × 10^{-4} and β(40°) = 4.3 × 10^{-12} m^{-1} s^{-1}.

Cruicial to the case of the beacons is whether the Rossby waves can propagate vertically from their tropospheric source, requiring that the vertical wavenumber m is real and non-zero (i.e., m² > 0), which implies that the quantity u - c must be less than some critical velocity for vertical propagation, U_c, given by (Sánchez-Lavega, 2011):

\[ U_c = \frac{\beta}{k^2 + f^2 + (f_0^2/N_b^2)} \]

(5)

Evaluation of this expression requires a selection of the relevant horizontal sizes. The zonal wavenumber, k = 2π/L_x, depends on the selected east–west dimension L_x – the width of B0 (some 70° at its largest), or the horizontal separation of B1 and B2 (extrapolated to be 39° in Section 3 at the time of the storm outbreak, but increasing as the two beacons moved apart). The meridional wavenumber, l = 2π/L_y, assumes a north–south distance of 30° latitude, with latitudinal extent of the beacon. Taking 39°–70° as a suitable range for L_y, we find critical velocities in the range U_c = 24–30 m/s. Even assuming a wave of 360° horizontal extent, the largest value we obtain for U_c is 32.5 m/s. To assess the vertical propagation, we must determine the ambient zonal velocity at the altitude of the beacon (see Appendix A), which is expected to vary with season. From Fig. 11d, the beacon is centred in a band where the 2-mbar velocities have been affected by the storm before August 2011. This would imply that u - c ≈ 25 m/s < U_c (as c ≈ - 25 m/s), satisfying the condition for the vertical propagation of Rossby wave disturbances of this horizontal size and drift rate.

However, we caution that the variation of the zonal wind with latitude on the giant planets implies that we should replace β with a modified value which includes the effects of the meridional curvature of the zonal winds (note that we do not account for vertical curvature of the winds in this calculation). If we replace β by an effective β_e = β - df/dy, we find that β_e is negative at 40°N because the second derivative of the zonal velocity field is positive. Hence the corresponding critical velocity U_c is retrograde and it is inconclusive as to whether planetary waves are able to propagate upwards from a tropospheric source. However, wave disturbances could still enter this evanescent region, their amplitudes decaying away from the source but potentially leaking into the upper atmosphere (e.g., Achterberg and Flasar, 1996). In summary, a beacon origin due to the vertical propagation of planetary waves from the tropospheric storm is supported by our simple calculations.
(in addition to gravity waves), but more sophisticated modelling is required to confirm this interpretation.

5.2. Conditions within the stratospheric anticyclone

At first glance, the thought of a warm stratospheric hotspot as an anticyclone may be confusing. For example, both Jupiter’s Great Red Spot (GRS) and the anticyclone formed in Saturn’s troposphere by the storm (Fletcher et al., 2011; Sánchez-Lavega et al., 2011) are cold-core features, with temperatures increasing radially outwards leading to windshears that decelerate their anticyclonic motion. On the other hand, Saturn’s stratospheric anticyclone has temperatures decreasing radially outwards, accelerating its clockwise motion with height. The tropospheric convective plumes served as a source of anticyclonic vorticity (implied by the upwelling and divergence of material on a rotating planet), and this anticyclonicity was somehow transported vertically throughout the atmosphere, leading to the clockwise rotation of the beacon at high altitude.

When the Rossby number is less than unity, angular momentum conservation on a rotating planet (the Coriolis effect) requires that high pressure systems are surrounded by clockwise winds to balance the pressure gradient in the northern hemisphere – an anticyclone. On Earth we think of anticyclones (high pressure regions) as regions of subsiding, dry air, the opposite of what we find for Jupiter’s GRS. In the free atmosphere, winds will flow parallel to isobars, but frictional forces on a solid surface causes air to cross isobars, leading to divergence at the base of a high-pressure system (i.e., the velocity has a component outwards from the centre). To conserve the mass of air within the system, air therefore sinks over high pressure systems on Earth. In Saturn’s troposphere, the frictional damping responsible for the deceleration of the zonal winds with altitude occur at the upper boundary of the vortices (due to some unidentified source of damping, e.g., Conrath and Pirraglia, 1983), rather than at the bottom, which may explain why anticyclones in giant planet tropospheres are regions of upwelling and divergence at the top. But what of anticyclones in the stratosphere? Here the situation may be reversed and more ‘Earth-like’, with friction at the lower boundary of the high-pressure system causing outflow at the bottom, and mass conservation resulting in atmospheric subsidence within the beacon.

The amount of energy required to heat up the stratospheric beacon can be estimated from the August 2011 mapping observation. We calculate the energy density required at each pressure level to produce the temperature difference $\Delta T(p)$ compared to the quiescent background:

$$e = \frac{1}{m g} \int_{p_0}^{p_{\text{max}}} C_p(p) \Delta T(p) dp$$

where $e$ is the energy density per unit area, $C_p(p)$ is the specific heat at constant pressure calculated for each pressure level, $\Delta T(p)$ is the temperature perturbation, $g$ is the local gravitational acceleration and $m$ is the mean molecular weight. The integration was performed between 0.1 and 100 mbar. Summing this within the oval determined by the peripheral vortex collar in Fig. 11, we estimate a total energy of $10^{22}$ J to produce this temperature perturbation, which is only a small fraction of (i) the $6.2 \times 10^{24}$ J of energy emitted by Saturn in a single year (Hanel et al., 1983), and (ii) the estimated total energy of $10^{24}$ J released by lightning in the thunder storm during its lifetime (Fischer et al., 2011).

6. Conclusions

The evolution of Saturn’s stratospheric vortex was described using spatial mapping of hydrocarbon emission and retrievals of atmospheric temperatures, stability, winds and composition. Table 5 provides an overview of the key events in the vortex evolution, and here we summarise some of the key findings of this study.

1. Thermal imaging and spectroscopy sensitive to hydrocarbon emission in Saturn’s stratosphere have revealed the never-before-seen stratospheric perturbations triggered by a planetary-scale tropospheric storm. The evolutionary sequence can be divided into three phases: (i) the formation and intensification of two distinct warm airmasses between January and April 2011; (ii) the merging and deceleration to form the large single beacon between April and June 2011; and (iii) the acceleration and slow cooling of the beacon since July 2011. One beacon (B1) was more extended, and located directly above the convecting storm head in the troposphere, moving west at 2.73 ± 0.1 deg/day. The second beacon (B2) was more compact and at a more southerly latitude, moving west at 0.6 ± 0.1 deg/day. This difference in drift rates caused the two beacons to encounter one another, when they behaved as two coherent anticyclonic vortices and merged in late April 2011, forming a single large beacon complex (B0).

2. Beacon B0 was tracked in Saturn’s stratosphere from May 2011 to March 2012, and persisted as the largest vortex detected in our Solar System to date (covering approximately 65° longitude between peripheral velocity peaks in August 2011, 50,000 km at 39°N, and extending between 20°N and 60°N), containing the hottest temperatures measured to date (the maximum temperature retrieved was 221.6 ± 1.4 K at 2 mbar on May 5th 2011). B0 moved rather slowly ($1.6 \pm 0.2$ deg/day) for the first two months post-merger, but then accelerated to 2.70 ± 0.04 deg/day ($-24.5 \pm 0.4$ m/s at 39.5°N) where it remained for the remainder of the observing period. B0 had no association with the tropospheric storm head after the merger, existing as an independent phenomenon, a ‘free mode’ of the stratospheric circulation. No perturbations to visible clouds and hazes were detected during this period. B0 was accompanied by a warm tail of stratospheric emission with peak temperatures between 0.3–0.8 mbar and 48–50°N, generating stratospheric thermal windshears to accelerate a retrograde jet at 47°N and a prograde jet at 52°N. The peak temperatures within the vortex are cooling slowly, by approximately 0.11 ± 0.01 K/day (40.2 ± 3.7 K/year, faster than the radiative time constant at 2 mbar for Saturn’s nominal atmosphere), and the width of the 160-K contours is shrinking by 0.16 ± 0.01 deg/day. Both of these processes, cooling and shrinking, will ultimately return Saturn’s northern stratosphere to quiescent conditions.

3. Retrievals of atmospheric temperatures show that the merger of B1 and B2 was associated with a downward shift in the peak temperatures from 0.5 mbar pre-merger to 2 mbar post-merger, a translocation of approximately 1.4 scale heights. Such a downward shift partially explains the enhancement in the acetylene abundance by a factor of 3 within the vortex, but fails to explain the lack of variability in stratospheric ethane. Despite the negative lapse rate introduced into the stratosphere by the presence of the beacon, calculations of the temperature gradient confirm that the atmosphere above the beacon remains sub-adiabatic. Although the sensitivity of the CIRS nadir spectra is minimal above the 0.1-mbar level, the retrievals do hint at structure associated with the stratospheric disturbance at even lower pressures, which could ultimately be explored by higher-resolution spectroscopy.

4. Thermal windshear calculations allow us to analyse the velocities of the vortex, confirming a Rossby number much smaller than unity (hence geostrophy is maintained), and demonstrating anticyclonic vorticity (clockwise rotation) with peripheral velocities of 200–400 m/s at 2 mbar. The clockwise peripheral velocities of the two original beacons was smaller (70–140 m/s).
The tropospheric storm, featuring upwelling and divergence, was a source of anticyclonic vorticity for the stratosphere.

5. Stratospheric acetylene at 2 mbar is uniformly enhanced within peripheral high velocity collar by a factor of three compared to the typical mean abundance of 0.4 ± 0.05 ppm at this latitude, varying around a mean of 1.3 ± 0.2 ppm, consistent with the downward displacement of the vortex post-merger. Ethane variability is within our retrieval uncertainty. The differences between the two hydrocarbons may be related to unusual chemistry within the hot vortex; differences in the vertical distribution; or possible high-altitude structure (sensed by C$_2$H$_2$ line cores) that cannot be reliably measured by CIRS.

6. The origins of the beacons remain uncertain, but numerical simulation shows that the tropospheric convective plumes impinging on the tropopause acted as strong source of waves (gravity waves, planetary waves), which extend in three dimensions away from the source. The upward transmissivity of the wave activity depends on the details of the temperature and wind structure near the tropopause, and simple calculations indicate that vertical Rossby wave propagation was permitted during this springtime epoch. The energy and momentum transported by these waves was deposited at stratospheric altitudes, and non-linear effects caused the formation of the two vortices, transferring energy from smaller to larger scales (an example of an inverse energy cascade).

The mystery of why this storm and other planetary-scale disturbances occur during Saturn's northern spring and summer remains unanswered. As seasonal changes to the vertical T(p) profile occur only at altitudes above the 500–700 mbar level, the seasonal triggering of the disturbance at great depth is rather puzzling. However, these seasonal changes may modify the circulation within the upper troposphere down to the topmost clouds, increasing the convective conditions and thereby altering the meridional circulation beneath the clouds, permitting large convective events from below (that are themselves independent of season) to extend higher than they could otherwise. This would suggest that the moist convective features could have been pre-existing (for example, Jupiter's moist convective events occur almost randomly in time), but only when seasonal thermal changes occurred did the full outbreak erupt.

This is the first time that the formation of a large stratospheric anticyclone, associated with a tropospheric convective storm, has been observed on a giant planet. Jupiter's north pole features a 'hot spot' at 60°N that is a region of high UV, infrared and X-ray emission (Gladstone et al., 2002, and references therein), but this is likely to be related to heating by auroral processes. Mid-latitude convective storms on Jupiter have been monitored in thermal-IR imaging (e.g., Sánchez-Lavega et al., 2008), but to date there have been no detections of stratospheric heating, making Saturn's stratospheric anticyclone unique. Stratospheric anticyclones are common features of the Earth's stratosphere, most common in northern winter and southern spring (Harvey and Hitchman, 1996; Harvey et al., 2004), and possibly associated with upward extension of planetary waves from tropospheric sources (e.g., Charney and Drazin, 1961). Their growth and intensification often accompanies the distortion and displacement of the polar cyclonic vortices from the poles (e.g., the Aleutian high displacing the Arctic vortex and the Australian high displacing the Antarctic vortex) and are associated with sudden stratospheric warming events (e.g., Andrews et al., 1987). On Earth, stratospheric anticyclones are regions of strong gradients in potential vorticity and trace species (e.g., Manney et al., 1995), just as Saturn's vortex appears to be entraining a region of elevated C$_2$H$_2$.

These thermal-IR observations demonstrate the strong coupling over hundreds of kilometres between the tropospheric outburst and the stratospheric thermal structure. It is not known whether previous saturnian storms produced similar stratospheric perturbations, or whether they will in future. Many of the physical processes revealed by this infrared study will require numerical modelling to explore in greater detail, such as those processes responsible for (i) transporting energy into the stratosphere from the tropospheric perturbation; (ii) forming two warm airmasses at 0.5 mbar that moved at different velocities; (iii) merging the vortices and translating the peak temperatures down by 1.7 scale heights; and (iv) maintaining B0 as a large anticyclone, stable against dissipative effects over long time periods. The ultimate fate of Saturn's unique stratospheric vortex remains to be seen, and new convective storms during Saturn's northern summer (starting in May 2017) will be intensively studied to determine whether this will phenomenon will happen again.

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Appendix A. Stratospheric zonal winds

Given the large number of mysteries surrounding this new stratospheric phenomenon, we investigated whether the presence of a long-lived feature in Saturn's stratosphere could have enabled a direct measurements of stratospheric winds. If the beacons can be truly considered as vortices or wave-like phenomena, their motions would be determined by the types of waves responsible and the background zonal velocity, $u$. However, the fact that B1 (and later B0) moved with the same velocity as the tropospheric storm head, despite a vertical separation of hundreds of kilometres, suggested a highly barotropic atmosphere, with little variation of the retrograde velocity with altitude. This is partially borne out by Fig. 1Id, which suggests zonal velocities near zero at 2 mbar in August 2011. However, the ambient zonal flow varies with time, which would have a significant effect on the trapping of planetary
waves in the troposphere (i.e., preventing the vertical propagation). If we had taken mid-latitude zonal winds from Cassini's prime mission (2004–2008 during northern winter), the 39.2°N jet, we find that the jet should switch from a retrograde velocity of ~20 m/s at 500 mbar, to zero velocity near 150 mbar, then increasing to form a prograde jet of ~100 m/s at 1 mbar, and planetary waves would have been trapped in the troposphere (u ≈ c ≈ 125 > Ut).

This large stratospheric velocity difference between wintertime and springtime mid-latitudes prompted us to explore the effects on the vertical shear on the zonal winds due to Saturn's seasonal asymmetry in temperatures (e.g., Friedson and Moses, 2012). Cassini observations since 2004 have revealed the slow variation of stratospheric temperatures (Fletcher et al., 2007, 2010), which is reproduced by numerical modelling of the radiative response to insolation changes. During southern summertime conditions at the start of the Cassini observations, the thermal gradient dT/ddy was largely negative, such that the northern winter hemisphere has a positive vertical shear, du/dz, accelerating the prograde stratospheric flow as described above. The southern summer hemisphere had a corresponding negative vertical shear, promoting retrograde flow in the southern stratosphere. As the seasons progress into northern spring and summer by May 2017, the thermal asymmetry should reverse to a positive dT/ddy, causing a negative vertical shear, du/dz, in the northern summer hemisphere. These simple arguments suggest that the direction of the stratospheric zonal flows are seasonally reversing, thus changing the transmissivity of the lower stratosphere to Rossby wave propagation.

We used the radiative climate model of Greathouse et al. (2008) (applied to Saturn by Fletcher et al. (2010)) to simulate the changing hemispherical stratospheric dT/ddy with time through an entire saturnian year. The magnitude of the calculated zonal velocity is highly sensitive to (i) the shielding effects of the rings (the model lacks advective motions, and therefore produces artificially large cooling effects in the ring shadowed regions); (ii) the meridional distribution of stratospheric hydrocarbons, particularly C2H2 and CH2=CH2, which are held constant in the model; (iii) the range of latitudes over which we average the zonal wind field; and (iv) the assumed level of zero motion, here arbitrarily set to the tropopause. The temperature predictions were used to calculate the vertical windshear with latitude and season, and then averaged over the 30–50°N range of the beacon. The model suggests that the magnitude of the prograde stratospheric flow has been decreasing at 40°N for the majority of the Cassini mission, and it is not inconceivable that retrograde motion could have been in place in the stratosphere when the storm first erupted, as suggested by Fig. 11d. Conditions may have been 'just right' for planetary wave propagation in Saturn's springtime to form the beacon. However, given the uncertainties associated with the extrapolation of the thermal winds above the cloud tops, we must await direct measurements of middle-atmospheric circulation to be certain of this temporal variability of stratospheric winds.

References