<table>
<thead>
<tr>
<th><strong>Manuscript Number:</strong></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Full Title:</strong></td>
<td>Enceladus as an active body</td>
</tr>
<tr>
<td><strong>Article Type:</strong></td>
<td>Regular review paper</td>
</tr>
<tr>
<td><strong>Keywords:</strong></td>
<td>Enceladus; Satellites; Saturn System</td>
</tr>
</tbody>
</table>
| **Corresponding Author:** | Juergen Schmidt  
   University of Oulu  
   Oulu, FINLAND |
| **Corresponding Author's Institution:** | University of Oulu |
| **Corresponding Author's Secondary Institution:** |  |
| **First Author:**      | Sascha Kempf |
| **First Author Secondary Information:** |  |
| **Order of Authors:**  | Sascha Kempf  
   Juergen Schmidt  
   T. Brockwell  
   Thomas Cravens  
   Larry W. Esposito  
   Katherina Fiege  
   Bernde Giese  
   Matthew M. Hedman  
   Paul Helfenstein  
   Terry A. Hurford  
   Ralf Jaumann  
   Geraint Jones  
   Sheila Kanani  
   Susan W. Kieffer  
   William Lewis  
   Brian A. Magee  
   Dennis L. Matson  
   Francis Nimmo  
   Robert Pappalardo  
   Mark E. Perry  
   Frank Postberg  
   Joachim Saur  
   John Spencer  
   Christophe Sotin |
Data from the Cassini-Huygens mission to Saturn revealed that the mid-sized icy satellite Enceladus is geologically active: A region around the south pole of Enceladus is anomalously warm and a mixture of gases, dominantly water vapor, is found to flow out there from a system of cracks in the ice crust. Micron sized ice particles, entrained in the gas, are ejected to space. It became clear, that Enceladus is the dominant source for magnetospheric plasma in the Saturn system, as well as for neutrals and dust, which forms the E ring of Saturn. What is so particular about Enceladus? Why is the activity localized at the south pole? How is the heat ultimately generated? How old is the phenomenon and how did it evolve? A complete, consistent answer to these questions is not known to date. Nevertheless, much progress has been made in analyzing and understanding various aspects of the activity, which gives important constraints and directions for future research. It is the purpose of this paper to review the current understanding of Enceladus’ geodynamics, geochemistry, and the interaction with the saturnian system.
Enceladus as an active body

Sascha Kempf · Jürgen Schmidt · T. Brockwell · Thomas Cravens · Larry W. Esposito · Katherina Fiege · Bernd Giese · Matthew M. Hedman · Paul Helfenstein · Terry A. Hurford · Ralf Jaumann · Geraint Jones · Sheila Kanani · Susan W. Kieffer · William Lewis · Brian A. Magee · Dennis L. Matson · Francis Nimmo · Robert Pappalardo · Mark E. Perry · Frank Postberg · Joachim Saur · John Spencer · Christophe Sotin · Frank Spahn · Benjamin D. Teolis · Gabriel Tobie · Hunter Waite · David T. Young · Mikhail Zolotov

The first two authors organized the workshop Active Enceladus at the International Space Science Institute (ISSI) in Bern, Switzerland, from which this paper results. All others are listed alphabetically.

Sascha Kempf
LASP, University of Colorado, 1234 Innovation Drive, Boulder, CO 80309-0392, USA

Jürgen Schmidt
University of Oulu, PL 3000, 90014 Oulu, Finland

T. Brockwell
Space Science and Engineering Division, Southwest Research Institute, San Antonio, TX 78238, USA

Thomas Cravens
University of Kansas, Lawrence, Kansas, USA

Larry W. Esposito
LASP, University of Colorado, 1234 Innovation Drive, Boulder, CO 80303-7814, USA

Katherina Fiege
IGEP, Universität Braunschweig, Mendelsohnstr. 3, 38106 Braunschweig, Germany

Bernd Giese
DLR, Institut für Planetenforschung, Planetengeologie, Rutherfordstraße 2, 12489 Berlin, Germany

Matthew M. Hedman
Cornell University, 322 Space Sciences Building, Ithaca, NY 14853-6801, USA

Paul Helfenstein
Center for Radiophysics and Space Research, Cornell University, Ithaca, NY 14853-6801, USA

Terry A. Hurford
Planetary Systems Laboratory, NASA Goddard Space Flight Center, Greenbelt, MD 20771, USA

Ralf Jaumann
DLR, Institut für Planetenforschung, Planetengeologie, Rutherfordstraße 2, 12489 Berlin, Germany

Geraint Jones
MSSL, University College London, Holmbury St Mary, Dorking, Surrey, RH5 6NT, UK

Sheila Kanani
MSSL, University College London, Holmbury St Mary, Dorking, Surrey, RH5 6NT, UK

Susan W. Kieffer
University of Illinois at Urbana-Champaign, 1301 W. Green St., Urbana, IL 61801, USA

William Lewis
Space Science and Engineering Division, Southwest Research Institute, San Antonio, TX 78238, USA
Abstract  Data from the Cassini-Huygens mission to Saturn revealed that the mid-sized icy satellite Enceladus is geologically active: A region around the south pole of Enceladus is anomalously warm and a mixture of gases, dominantly water vapor, is found to flow out there from a system of cracks in the ice crust. Micron sized ice particles, entrained in the gas, are ejected to space. It became clear, that Enceladus is the dominant source for magnetospheric plasma in the Saturn system, as well as for neutrals and dust, which forms the E ring of Saturn. What is so particular about Enceladus? Why is the activity localized at the south pole? How is the heat ultimately generated? How old is the phenomenon and how did it evolve? A complete, consistent answer to these questions is not known to date. Nevertheless, much progress has been made in analyzing and understanding various aspects of the activity, which gives important constraints and directions for future research. It is the purpose of this paper to review the current understanding of Enceladus’ geodynamics, geochemistry, and the interaction with the saturnian system.
1 Introduction

The full extent of Enceladus’s activity was ultimately discovered in Cassini data. But already much earlier its remarkable surface properties, seen in Voyager images, led to the conclusion on geophysical activity (Squyres et al. 1983; Haff et al. 1983), with the compelling suggestion that cryo-volcanic eruptions on Enceladus are the source of the dusty E ring (Pang et al. 1984; Showalter et al. 1991a). That ring was discovered by Feibelman (1967b) in images taken from earth, and earth-bound observations at edge-on configurations showed that the ring is densest around the orbit of Enceladus (Baum et al. 1981), while the flybys of the Pioneer and Voyager spacecraft had confirmed the close relation of the E ring and Enceladus.

In a detailed analysis of Voyager images Kargel and Pozio (1996) studied the surface of Enceladus, assigning probable ages to various surface units from the crater records (see also Kargel (2006)). They found very old terrains, that might be nearly as old as the Saturn system, and regions practically void of craters, likely not older than \(10^7 - 10^8\) years. They concluded that Enceladus must have undergone multiple episodes of activity over a time span of billions of years, leading to a large scale resurfacing, requiring large amounts of tidal heating, probably exceeding the current steady state value. Remarkably, Showalter et al. (1991a) interpreted the narrow size distribution of the E ring, inferred photometrically as a strong support for the hypothesis of geyser like cryo-volcanic ejection of ice grains.

In Cassini data the first indication for Enceladus’ geophysical activity was found in perturbations of Saturn’s magnetic field near the moon. The Cassini magnetometer team (Dougherty 2005) interpreted this observation in terms of a partial atmosphere of Enceladus in the southern hemisphere, decelerating and deflecting the corotational plasma, leading to the observed perturbation of the planetary magnetic field.

In the subsequent targeted flyby E2 (Tab. 1) the intense geophysical activity became evident in the data returned by the Cassini instruments, including the discovery of the hot south polar region (Spencer et al. 2006), associated with the system of cracks dubbed ‘Tiger Stripes’ (Porco et al. 2006), the peculiar diversity of surface properties (Porco et al. 2006) and its composition (Brown et al. 2006), the vapor plume emerging from the Tiger Stripes (Waite et al. 2006; Hansen et al. 2006), the dust plume (Spahn et al. 2006b; Porco et al. 2006) formed by micron sized water ice grains, and the plumes’ interaction with the Kronian magnetosphere (Dougherty et al. 2006; Kivelson 2006; Tokar et al. 2006; Jones et al. 2006) and the E ring (Spahn et al. 2006b).

In the following years Enceladus attracted the attention of researchers from a large variety of scientific disciplines. This paper gives a summary of progress achieved and it is organized as follows. In Section 2 we cover various aspects of Enceladus’ geodynamics, as its shape, tectonic structure and craters, internal structure, tidal heating and heat transfer, internal dynamics and ocean evolution, as well as the coupled orbital and thermal evolution with the possibility of a recurrent transient activity. In Section 3 we summarize the constraints on the geochemistry, composition of rocky core and ice shell, indications on aqueous and dry processes, as well as plausible scenarios of chemical evolution. Section 4 is devoted to the Enceladus plume, constraints on its gas and dust components, and a discussion of scenarios for plume formation. Finally, in Section 5 the effect on the Saturnian system is discussed, the replenishment of Kronian plasma, the neutral torus, and the dusty E ring by material from Enceladus, the interaction with the magnetosphere, and delivery of species to the outer magnetosphere.
Enceladus as an active body

<table>
<thead>
<tr>
<th>Encounter</th>
<th>Orbit</th>
<th>Date</th>
<th>Closest Approach</th>
<th>Geometry</th>
<th>Plume Distance</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Time (UT)</td>
<td>Altitude (km)</td>
<td>Speed (km s(^{-1}))</td>
</tr>
<tr>
<td>E0</td>
<td>3</td>
<td>2005-02-17</td>
<td>03:30</td>
<td>1260</td>
<td>6.6 U</td>
</tr>
<tr>
<td>E1</td>
<td>4</td>
<td>2005-03-09</td>
<td>09:08</td>
<td>497</td>
<td>6.6 U</td>
</tr>
<tr>
<td>E2</td>
<td>11</td>
<td>2005-07-14</td>
<td>19:55</td>
<td>166</td>
<td>8.2 U</td>
</tr>
<tr>
<td>E3</td>
<td>61</td>
<td>2008-03-12</td>
<td>19:06</td>
<td>48</td>
<td>14.4 P</td>
</tr>
<tr>
<td>E4</td>
<td>80</td>
<td>2008-08-11</td>
<td>21:06</td>
<td>49</td>
<td>17.7 P</td>
</tr>
<tr>
<td>E5</td>
<td>88</td>
<td>2008-10-09</td>
<td>19:06</td>
<td>25</td>
<td>17.7 P</td>
</tr>
<tr>
<td>E6</td>
<td>91</td>
<td>2008-10-31</td>
<td>17:14</td>
<td>169</td>
<td>17.7 P</td>
</tr>
<tr>
<td>E7</td>
<td>120</td>
<td>2009-11-02</td>
<td>07:42</td>
<td>99</td>
<td>7.7 S</td>
</tr>
<tr>
<td>E8</td>
<td>121</td>
<td>2009-11-21</td>
<td>02:10</td>
<td>1697</td>
<td>7.7 S</td>
</tr>
<tr>
<td>E9</td>
<td>130</td>
<td>2010-04-28</td>
<td>00:10</td>
<td>100</td>
<td>6.5 S</td>
</tr>
<tr>
<td>E10</td>
<td>131</td>
<td>2010-05-18</td>
<td>06:56</td>
<td>437</td>
<td>6.5 U</td>
</tr>
<tr>
<td>E11</td>
<td>136</td>
<td>2010-08-13</td>
<td>22:31</td>
<td>2555</td>
<td>6.8 S</td>
</tr>
<tr>
<td>E12</td>
<td>141</td>
<td>2010-11-30</td>
<td>11:54</td>
<td>46</td>
<td>6.3 N</td>
</tr>
<tr>
<td>E13</td>
<td>142</td>
<td>2010-12-21</td>
<td>01:08</td>
<td>48</td>
<td>6.2 N</td>
</tr>
<tr>
<td>E14</td>
<td>154</td>
<td>2011-10-01</td>
<td>13:52</td>
<td>99</td>
<td>7.4 P</td>
</tr>
<tr>
<td>E15</td>
<td>155</td>
<td>2011-10-19</td>
<td>09:12</td>
<td>1231</td>
<td>7.4 P</td>
</tr>
<tr>
<td>E16</td>
<td>156</td>
<td>2011-11-06</td>
<td>04:59</td>
<td>497</td>
<td>7.4</td>
</tr>
<tr>
<td>E17</td>
<td>163</td>
<td>2012-03-27</td>
<td>18:30</td>
<td>74</td>
<td>7.5 S</td>
</tr>
<tr>
<td>E18</td>
<td>164</td>
<td>2012-04-14</td>
<td>14:02</td>
<td>74</td>
<td>7.5 S</td>
</tr>
<tr>
<td>E19</td>
<td>165</td>
<td>2012-05-02</td>
<td>09:31</td>
<td>74</td>
<td>7.5 S</td>
</tr>
<tr>
<td>E20</td>
<td>223</td>
<td>2015-10-14</td>
<td>10:41</td>
<td>1839</td>
<td>8.5 S</td>
</tr>
<tr>
<td>E21</td>
<td>224</td>
<td>2015-10-28</td>
<td>15:23</td>
<td>49</td>
<td>8.5 S</td>
</tr>
<tr>
<td>E22</td>
<td>228</td>
<td>2015-12-19</td>
<td>17:49</td>
<td>4999</td>
<td>9.5 S</td>
</tr>
</tbody>
</table>

Table 1 Parameters of the Cassini orbiter encounters with Enceladus. The last three flybys E20, E21, and E22 are listed here as planned for the Cassini Solstice Mission as of this writing. The plume source is here assumed as the south pole of Enceladus' surface. The flyby geometries are defined as follows: U = upstream; P = North- South plume; S = southern; N = northern.

2 Geodynamics

2.1 Observations

2.1.1 Global shape and topography

The topography of Enceladus is not properly known yet. Topography requires an equipotential surface for referencing heights, but elevation models of Enceladus presented so far (Schenk and McKinnon 2009; Giese and The Cassini Imaging Team 2010) refer to the best-fit ellipsoid (256.6 × 251.4 × 248.3 km) which, however, is non-hydrostatic (Thomas et al 2007) and thus does not form an equipotential surface. As a consequence, large-scale features may actually be smoother when referenced to the true equipotential surface (which is currently unknown). Schenk and McKinnon (2009) have studied whether a negative core density anomaly could explain the observed 100 km wavelength depressions simply by deflection of the equipotential surface, and concluded that this explanation did not work. The depressions are more likely due to ice shell density and/or thickness variations (Schenk and McKinnon 2009). However, it is unclear how a convecting ice shell could sustain such lateral variations on the long term, and indeed the ice shell might well be static (conductive) in most places (see section 2.3.2). Further modeling is needed here.

Currently existing elevation models (Schenk and McKinnon 2009) Giese and The Cassini Imaging Team 2010 cover about 60% of Enceladus’ surface. They show several up to 2 km deep depressions comparable in size to Enceladus’ radius. The most pronounced one (centered at 10°S/230°E) is ovoid-shaped (200 × 140 km) and comprises both older cratered terrains and younger, resurfaced terrains implying that it does not correlate with geologic boundaries (Schenk and McKinnon 2009). Likewise, there is a more circular 250 km depression (centered at 10°N/309°E) with a distinct lower-lying area (70 × 120 km) not correlating with geologic boundaries. The South Polar Terrains (latitudes > 65°) correlate with a ~ 700 deep depression (Thomas et al 2007). There also exist large-scale convex-shaped features on Enceladus (71°N/38°E and 30°S/300°E) (Giese and The Cassini Imaging Team 2010).
These 200 km bulges are about 1 km high and, contrary to the depressions, correlate with resurfacing features. Local, small-scale features on Enceladus reach elevation similar to those at large scales: (i) 15-20 km craters reach depths of up to 1.8 km, (ii) the central peak of Aladdin crater (63°N/341°E, D∼3 km) stands about 1 km above the local surroundings, (iii) the Harran Sulci (3°N, 130°E), an 11 km wide half-graben structure, has an elevated flanking ridge up to 1 km high (Giese et al 2008), (vi) an active 2 km wide segment in Enceladus’ South Polar Terrains forms a V-shaped trough up to 500 m deep (Giese and The Cassini Imaging Team 2010).

2.1.2 Thermal emission

The total endogenic power of Enceladus’ south polar terrain has been recently reevaluated using the 2008 Cassini Composite Infrared Spectrometer (CIRS) 10 to 600 cm$^{-1}$ spectra (Howett et al 2011). The new estimate of $15.8 \pm 3.1$ GW is about three times higher than the previous estimate of $5.8 \pm 1.9$ GW (Spencer et al 2006), which was derived from $> 600$ cm$^{-1}$ observations, because the new estimate includes low-temperature emission that the earlier higher-wavenumber observations were not sensitive to. This indicates that Enceladus’s south polar terrain is even more thermally active than initially thought.

The latest CIRS > 600 cm$^{-1}$ observations confirm that the maximum of emission occurs along the tiger stripes and most of the emissions are confined to a narrow, intense region no more than a kilometer (half a mile) wide along the faults. Emission is continuous along the tiger stripes at 5 -10 km resolution (Howett et al 2011), but shows large spatial variations along-strike at higher resolution (Spencer et al 2011). The peak temperatures observed by CIRS along Damascus and Baghdad Sulci, the most active tiger stripes, are estimated to about 190 K, over widths if tens of meters along the stripes, but higher temperatures are possible over narrower regions closer to the fractures.

2.1.3 Surface physical properties

Enceladus’ surface is mostly composed of pure water ice, amorphous and crystalline, except near its south pole where light organics and CO$_2$ have also been identified, particularly in the Damascus, Baghdad, Cairo and Alexandria Sulcus regions (Brown et al 2006). Variations in the main water ice absorptions at 1.04, 1.25, 1.5, and 2.0 µm indicate that the averaged particle size varies on Enceladus’ surface and is directly correlated with geological units (Jaumann et al 2008). On a regional scale heavily cratered terrains exhibit a mean particle size of 15 ± 5 µm, larger particles with diameters up to 30 ± 10 µm are correlated either with fresh impact crater ejecta or with ridges and fractures marking the transition zone to the lightly cratered plains. Tectonically deformed regions exhibit particles ranging from 40 µm about 120 µm with the largest ones (about 2% reaching values of about 0.2 mm) located close to the center of the south polar sulci (Brown et al 2006; Jaumann et al 2008; Spencer et al 2009).

Similar patterns are seen in the ISS multispectral images of the south polar region, which show enhanced near-infrared absorption near the tiger stripes and other tectonically disrupted regions, consistent with larger grain sizes (up to100 µm) near the sulci (Porco et al 2006). Possible causes for the large grain sizes near the south polar sulci include preferential fallout of larger, slower, plume grains near their sources (Jaumann et al 2008; Hedman et al 2009), or sintering and growth of grains resulting from deposition of plume gases or from the enhanced temperatures near the south polar tectonic features (Spencer et al 2009). On a global scale, the visible orbital phase curve of Enceladus shows that the trailing hemisphere is brighter than the leading hemisphere (Buratti and Veverka 1984; Buratti et al 1998), and ground-based spectra show smaller ice particles on the trailing hemisphere (Verbiscer et al 2006). These variations may reflect hemispheric differences in Enceladus’ interaction with the E-ring.
Disk-integrated spectra of Enceladus indicate that the water ice in the equatorial and mid- to high-latitude regions is primarily crystalline (Grundy et al. 1999; Emery et al. 2005). However, Cassini spectra has detected locally abundant amorphous ice in the south polar region between the sulci, while the sulci themselves are dominated by crystalline ice (Brown et al. 2006; Newman et al. 2008). Amorphous water ice forms when it is condensed directly from the vapour to a solid at temperatures below about 100K (Brown et al. 2006; Newman et al. 2008). Away from the sulci, the surface temperatures are well below 100 K (Spencer et al. 2006), so amorphous ice should be able to form and remain stable for long periods of time there.

The icy surface of Enceladus is affected by various modification processes, including charged particle bombardment (implantation or sputtering), E-ring grain bombardment or coating, thermal processing, UV photolysis, and micrometeoroid bombardment. Water ions are the dominant sputtering agents, and rates are estimated to be about 0.1µm per year, or 10cm per million years (Johnson et al. 2008). The high albedo (Squyres et al. 1983; Verbiscer et al. 2005; Spencer et al. 2006) is probably due to a coating of plume particles, derived both directly from plume fallout and from re-impact of plume particles stored temporarily in the E-ring (Spahn et al. 2006b; Kempf et al. 2008; Spencer et al. 2009; Jaumann et al. 2009), suggesting a domination of coating over micrometeorite gardening. On the other hand, IR-spectra reveal a consistent pattern of increasing grain size mostly at tectonic features with decreasing surface age (Jaumann et al. 2008), which suggest a gradual comminution of particles over time by impact gardening or sputtering. This might be due to either uneven coating at steeper slopes or mass wasting in those areas.

2.1.4 Tectonic structures and stratigraphic map

Enceladus’ surface reveals a wide variety of icy tectonic structures that record a long history of tectonic deformation. These structures include rifts and tension cracks (fossae), lanes of striated and grooved terrains (sulci), networks of narrow ridges and domelike structures (dorsa), icy massifs and mountain chains, and peculiar quasi-parallel systems of parallel ridges that are most often associated with active south polar tectonism and volcanism. Slicing through nearly all of these features are ubiquitous sub-parallel arrangements of gossamer cracks with typical spacing of hundreds of meters (Miller et al. 2007; Barnash et al. 2006). Spencer et al. (2009) provided a detailed review of global tectonic structures that were known at the time. Since then, new Cassini imaging coverage has revealed previously unknown details of leading hemisphere and South Polar terrains, and new progress has been made in tectonic mapping and modeling of already-recognized features.

Detailed stratigraphic mapping (Fig. 1) of the South Polar Terrain region has been undertaken by Patthoff and Kattenhorn (2009, 2010, 2011). They distinguished between present day active tiger stripes and three earlier systems of paleo-tiger stripes, each system of which share a common orientation and relative age. The old tiger stripes are crosscut and offset by progressively younger fracture sets. Incremental changes in orientation of these systems suggest that Enceladus lithosphere has experienced non-synchronous rotation over time. The differing orientations of the modern and paleo-tiger stripe systems, respectively were found to correlate with corresponding fracture orientations in the deformation belts that separate the SPT province from the older terrains to the north. These correlations are consistent with hypothesis that tectonic spreading related to tiger stripes (Helfenstein et al. 2008a; Spencer et al. 2009; McKinnon and Schenk 2009; Barr 2008; Barr and Preuss 2010) has occurred over time and that corresponding convergence is manifested as fold-like belts and complex fracture systems at the boundary of the SPT (Helfenstein et al. 2006; Porco et al. 2006; Schenk and McKinnon 2009).

The most prominent ancient tectonic features on Enceladus are likely found within tectonized plains on the leading hemisphere and the trailing hemisphere.
Tectonic overprinting and evidence for early viscous relaxation is widespread and often only partial relict features may be recognized. As noted earlier, the broad hemisphere-centered regions of tectonic resurfacing are subdivided into smaller quasi-polygonal provinces that exhibit distinct tectonic styles and differing ages. On the leading hemisphere is a conspicuous province near 30°N, 90°W of curvilinear massifs and roughly orthogonal-trending wide, shallow troughs (cf. Pappalardo and Crow-Willard 2010) that define a crudely radial and concentric pattern relative to a point near 25°N, 125°W. This angular sector, about 65° in width, may be either the partial remains of an ancient impact basin with a diameter of about 180 km or perhaps a surface expression of an ancient, large diapir. These hypotheses are also consistent with the presence on the leading hemisphere of numerous subdued-relief circular albedo patterns and smooth circular patches up to 40 km in diameter (cf. Spencer et al 2009, Fig. 21.7). The peculiar quasi-radial wide-shallow troughs resemble extinct, topographically degraded examples of tiger stripes seen near the South Pole. While these features may have a different fracture origin from tiger stripes, their comparable morphology suggests that long ago they may have expressed a similar style of fissure volcanism (Helfenstein et al 2010b,a).

A peculiar system of rounded, ropy ridges separate the tiger stripe fractures in the SPT region. They are curved, quasi-parallel knobby ridges and clustered arrangements of elongated barrel-shaped knobs that were first recognized as grooved bands by Porco et al (2006), but are called funiscular plains (Helfenstein et al 2008b,a; Spencer et al 2009; Helfenstein et al ???a) and South Polar folds (Barr and Preuss 2010) in more recent works. They trend approximately parallel to...
tiger stripes and their lengths and kilometer-scale widths are comparable to those
nearby tiger stripe flanks. Near their distal ends where funisicular plains intersperse
appear to break into a more chaotic arrangement of elongated to equant knobs and
cylindrical segments. The superficial resemblance of offset patterns at the edges
of tiger stripes to terrestrial transform faults and stepovers, the morphological
similarity of an axial discontinuity along the Damascus Sulcus tiger stripe to ter-
restrial offset spreading centers and a plausible preliminary tectonic reconstruction
of South Polar paleo-terrains, suggest that the ropy ridges may form as folds due to
tectonic spreading around active tiger stripes, very crudely analogous to terrestrial
sea floor spreading (Helfenstein et al 2008a, ????b). Spencer et al. 2009 McKimmon
and Schenk 2009). This interpretation is supported by the atypical regional topo-
graphic flatness of the active South Polar Terrain (McKinnon and Schenk 2009), as
well as models of mobile-lid convection (Barr 2008 O’Neill and Nimmo 2010) and
terrestrial lava flow folding (Barr and Preuss 2010). The latter found that ropy folds
due to spreading could form in an ice shell with an upper high-viscosity boundary
layer of thickness < 400 meters, with a driving stress of 40 to 80 kPa, and strain
rate between 10−14 s−1 and 10−12 s−1. Their results suggest that resurfacing of
the SPT could occur over 0.05 to 5 Myr, consistent with its estimated surface age
from cratering statistics. A stratigraphically much older region containing rounded
ropy patterns of ridges that are similar in morphology and scale to those in the
SPT have recently been found on the leading hemisphere of Enceladus, near 10°S,
60°W, suggesting that perhaps tectonic spreading may have occurred elsewhere on
Enceladus at an earlier era. However, the overall physical arrangement of tectonic
features on the leading hemisphere significantly differs from the distinct pattern
seen in the SPT, so that it is not likely to simply be an early clone of the SPT
system (Pappalardo et al 2010), Helfenstein et al. (2010a).

Dorsa are straight to curvilinear narrow ridges hundreds of meters wide, with
rounded sub-kilometer scale relief. They are found within all major tectonized
provinces, but the best studied examples are Cufa Dorsa (3.2°N, 286.2°W) and
Ebony Dorsum (5.7°N, 280.5°W). Their lengths generally do not exceed a few to
tens of kilometers. They occur as isolated features or in crudely polygonal arrays
along with elongated domelike features with similar height and width dimensions.
Prominent examples on Enceladus trailing hemisphere may have originated as
thrust faults (Pappalardo and Crow-Willard 2010). However, some dorsum vari-
ants, specially near the South Polar Terrain, appear to arise along pre-existing
fractures, suggesting that they were extruded (Porco et al. 2006) or that they
may be transpressional features analogous to terrestrial positive flower structures
(Helfenstein et al. 2008b, ????a). This is especially true of a special class of narrow
dorsa, called shark fins that are found medially within tiger stripe fractures.

2.1.5 Crater distribution, crater morphology

Crater counts on Enceladus confirm the wide range of surface ages (Plescia and
and Schenk 2009). Depending on the flux model used, which can be either lunar-
like and therefore asteroidal (Neukum 1985) or, more plausibly, based on models of
cometary impact rates (Zahnle et al 2003), cratered plains materials have absolute
ages as old as 4.2 billion or 1.7 billion years, respectively. In general, the eastern
fractured terrain has formed either between 3.7 and 1.0 billion years ago in the
lunar-like scenario or it is as young as 200 to 10 million years based on the Zahnle
et al. model. The south polar terrain shows crater ages younger than 100 million
to few hundreds of thousands years, respectively, and consistent with the ongoing
activity at the tiger stripes.

Enceladus’ cratered plains distribution is unique in that it appears to have a
relative deficiency of craters for diameters ≤ 2 km and ≥ 6 km compared to the
other satellites heavily cratered plains (Kirchhoff and Schenk 2009). The data re-
ported by Kirchhoff and Schenk (2009) also indicates that the impact crater density
within the cratered plains changes with latitude. Specifically, both the north and
south mid-latitude regions have approximately three times higher density than
the equatorial region. The “missing” small and large craters in Enceladus cratered plains may be due to a combination of viscous relaxation of the larger craters, and burial of the relaxed large craters and small craters by south polar plume and possibly E-ring material. A large number of craters appears to viscously relaxed in the apparently young regions of the anti-Saturnian and trailing hemispheres, as well as in the older, upper northern latitudes (Bray et al. 2007; Smith et al. 2007; Smith 2008). The occurrence of viscous relaxation in various terrains suggest large geographical and time variations in heat flow.

2.2 Present day state

2.2.1 Tectonic structures, tidally driven activities

One prominent tiger stripe near Enceladus’ south pole consists of arcuate segments, resembling the shape of cycloidal cracks on Europa (c.f. Hurford et al. 2007). On Europa, these distinctive tectonic patterns were likely produced by periodic tidal stresses (Hoppa et al. 1999). The similarity between Enceladus tiger stripes and Europa’s cycloidal ridges is in shape only; they are quite distinct from one another morphologically. Nevertheless, the similar shape of the tiger stripe suggests that its formation may have been similarly controlled by diurnal tidal stresses (Hurford et al. 2007; Nimmo et al. 2007). If this is the case, diurnal tidal stresses may play an important role, driving activity on Enceladus.

Diurnal tidal stress on Enceladus is a natural product of its orbital eccentricity. Enceladus’ finite orbital eccentricity causes small daily changes in the distance between Enceladus and Saturn, affecting the height of the tide raised on the satellite by the planet. Besides affecting the height of the tide on Enceladus, the orbital eccentricity also causes the longitude of the tidal bulge to oscillate as it tracks the position of Saturn throughout an orbit. This constant reshaping of Enceladus’ shape produces stress on the surface. However, the tidal deformation and hence the resulting stress get significant only if a liquid water layer decouples the ice shell from the rocky core (Hurford et al. 2007; Nimmo et al. 2007; Tobie et al. 2008). Olgin et al. (2011) recently showed that tidal stress can reach values of 1 bar for ice shell thinner than 20 km, and that the stress could exceed the Coulomb criterion for shear failure along the tiger stripes if the ice shell thickness is less than 40 km. However, although tidal stress may be sufficient to re-activate pre-existing faults, it is probably insufficient to generate new faults.

Other source of stresses must be taken into account in order to explain the fracturing of the upper lithosphere. Change of volume due to thermal expansion/contraction or melting/crystallization of the internal ocean can lead for instance to stresses exceeding several hundred of bars (Nimmo 2004; Manga and Wang 2007; Mitri and Showman 2008). Changes in the solid surface position with respect to the tidal bulge can also result in significant stresses. Non-synchronous rotation (NSR) involves longitudinal shifts in position, while reorientation of the rotational axis, called “true polar wander”, also includes latitudinal shifts. Both mechanisms result in predictable global stress patterns (e.g. Melosh 1980; Leith and McKinnon 1996). Such a reorientation of the rotational axis results from perturbation of the moment of inertia of the body owing to large-scale internal or surface loads, and it is also thought to have a play role in the evolution of Enceladus (Nimmo and Pappalardo 2006).

Once the lithosphere is pre-fractured, tidal stress is likely to be the main driver of the propagation and persistence of cracks, by periodically reactivating them. Although the amplitude of stress is relatively small, the energy associated with tidal motions is quite large and contributes significantly to the dynamics of the lithosphere. The Tiger Stripes in the south polar region exhibit higher temperatures than the surrounding terrain and are the sources of the observed eruptions (Porco et al. 2006; Spencer et al. 2006; Spitale and Porco 2007). Over Enceladus’ orbital period, stresses across the tiger stripes alternate from compressive to tensile, possibly allowing the faults to open, which could expose a subsurface volatile reservoir,
creating an eruption (Hurford et al. 2007). In a complementary process, periodic shear stress along the rifts may generate heat along their lengths (Nimmo et al. 2007) (see section 2.2.3). Model calculations of tidal shear stress along rifts in the south polar region of Enceladus shows that there is a reasonable correlation between observed hotspot locations and predictions of regions, which show high amounts of tidal shear (Nimmo et al. 2007; Smith-Konter and Pappalardo 2008).

2.2.2 Possible internal structures and thermal state

The bulk density of Enceladus suggests an interior with roughly equal masses of ice and silicate (Schubert et al. 2007). Because the shape of Enceladus is not close to that expected for a body in hydrostatic equilibrium \((a-c)/(b-c) \neq 4\), (Thomas et al. 2007), see Section 2.1.1, neither shape nor gravity measurements can be used to infer the moment of inertia of the body. Nonetheless, given the high present-day heat flux and indications of subsurface liquid water (see below), a differentiated structure is generally assumed (Schubert et al. 2007; Spencer et al. 2009) (Figure 2).

The strongest argument for the presence of liquid water beneath the surface of Enceladus is the detection of salty grains in the E-ring (Postberg et al. 2009) (see Section 3.1.3). Liquid at any location in the ice shell would tend to drain downward, forming a local sea or a global ocean. We view the former of these possibilities as more likely (see Section 2.3.3). A hot silicate core has been argued based on the need to melt the ice shell beneath the south pole (Collins and Goodman 2007) and to explain the observed NH\(_3\) in the plumes (Matson et al. 2007). Neither argument is strong: melting within the ice shell is a viable alternative (Tobie et al. 2008) and other pathways to generating NH\(_3\) have been suggested (Glein et al. 2008). The core of Enceladus is so small that any primordial heat will have been long lost via...
Fig. 3 Three examples of heat production by tidal friction: a) Dissipation of eccentricity tides associated with strong tidal flow in an thin and shallow internal ocean (Tyler 2011), b) volumetrically-distributed viscous dissipation in a convective ice shell (after Behounková et al 2010), c) shear heating associated with tidal motions along tiger stripes (Nimmo et al 2007).

conduction, and it is very unlikely that the core is sufficiently ductile that tidal heating is important (Roberts and Nimmo 2008; Tobie et al 2008).

2.2.3 Heat generation by tidal friction and orbital equilibrium

The high south polar heat fluxes suggest either correspondingly high present-day heat production rates, or some mechanism of storing heat and then releasing it rapidly. The issue of heat transport is addressed below (Section 2.2.4); here we focus on heat production.

Three main mechanisms of heat production have been proposed. A first one is dissipation associated with strong tidal flow in an internal ocean (Tyler 2009, 2011). Large dissipation might be obtained for a thick ocean (1-50 km) but required a non-zero obliquity. The obliquity is currently unknown, and it is expected to be rather small. Enceladus is likely to be in a Cassini state, where the obliquity should be equal or smaller than the orbital inclination (≤ 0.01°). For such a low obliquity value, obliquity tides remains unimportant (Tyler 2009, 2011). Tyler (2011) also predicts that large dissipation associated with eccentricity tidal forces might also be obtained but only for very thin oceans (≤ 360m, Fig. 3a). However, as Tyler’s model assume that the ice lithosphere is thin enough to not affect the ocean tidal response, his calculations are therefore only valid for shallow oceans (few hundred meters under the surface) which is incompatible with topographic evidence of bulge and depression on the order of 1-2 km, as well as with our understanding of Enceladus’ interior structure. Such a thin, shallow ocean would be above an less dense ice layer, which is gravitationally unstable.

A second dissipation mechanism is conventional tidal heating, involving volumetrically-distributed viscous deformation in response to the time-varying tidal potential (e.g. Tobie et al 2005; Mitri and Showman 2008a; Behounková et al 2010). As this dissipation mechanism depends on the viscosity (or temperature), it strongly varies within a convective ice shell and therefore has a strong impact on the dynamics of thermal instabilities (Fig. 3b) (Behounková et al 2010).

The third mechanism is localized, shear heating on faults, again driven by tidal deformation (Fig. 3c) (Nimmo et al 2007). In this case, the dissipation process is also associated with viscous friction, but the tidal deformation is controlled by strike-slip motions along the faults rather than volumetrically-distributed viscous deformation.

In the ice shell, the two latter processes are probably operating simultaneously. Like the ocean flow mechanism, these two solid-state mechanisms require the presence of a subsurface ocean, at least in the southern hemisphere: an ice shell
Enceladus as an active body

coupled directly to the silicate mantle does not deform sufficiently to generate the observed heating rates. Both mechanisms have difficulty explaining the degree-one pattern of heating, because tidal deformation is a degree-two feature i.e. hot spots should be observed at both poles. The most likely explanation is probably some kind of pre-existing thermal anomaly, which is not terribly satisfying.

Although the details of the process need further study, shear heating appears capable of explaining the observed south polar heat and vapour fluxes for reasonable parameter choices (Nimmo et al. 2007). Generating the required power in a localized warm diapir, either with a global ocean or a regional sea, is also possible as long as the viscosity is relatively low ($\sim 10^{13} - 10^{14}$ Pas or less) (Tobie et al. 2008). Dissipation in the ocean might also generate several gigawatts, but it requires very specific conditions in term of obliquity or ocean thickness, which could be met only during relatively short periods of time.

Irrespective of the mechanism of heat production, an extremely important point is that the observed current heat loss greatly exceeds the steady-state tidal heat production rate (Meyer and Wisdom 2007). This steady-state value arises because of the competition between the Enceladus:Dione 2:1 resonance and dissipation in Saturn, which increases the eccentricity e of Enceladus, and dissipation within Enceladus, which damps it. The steady state heat production rate depends on dissipation within Saturn. For a dissipation factor estimated to 18 000 in Saturn (e.g. Meyer and Wisdom 2007), Enceladus is currently losing heat at a rate faster than the long-term rate at which it can be produced, estimated to be 1.1 GW; further discussion of this issue is deferred to Section 2.3.4.

Physical libration and forced libration associated with secondary spin-orbit resonance (Wisdom 2004; Hurford et al. 2009) may also lead to enhanced tidal deformation and hence strong tidal heating. Hurford et al. (2009) showed, for instance, that for the case of a physical libration at the Cassini’s limit of $L = 1.5^\circ$, the tidal heating is potentially enhanced by $\sim 4.9$ times. Current efforts to directly detect Enceladus’ physical libration from Cassini imaging limits its amplitude to $\leq 15^\circ$ (6.6 km) (Porco et al. 2006). Rotational models of Enceladus with no global ocean predict that the librations related to the perturbations of the orbit by Dione may reach amplitude of the order of 1 km (Rambaux et al. 2010), significantly smaller than the value assumed by Hurford et al. (2009). There is no indication that enhanced dissipation associated with libration play a key role at present, but it is possible that Enceladus passed through such a high dissipative stage in a recent past, especially if it enters a secondary spin-orbit resonance (Wisdom 2004).

2.2.4 Heat transfer from deep interiors to the surface

The internal heat, mainly produced by tidal dissipation in the H$_2$O layer, may be transferred to the surface from the source region by conduction through the ice, or by solid-state convection of water ice, or by flow of liquid water or by circulation of water vapor. All these heat transport mechanisms probably contribute to the total observed thermal emission, but their relative contribution to the total heat transport strongly depend on the thermal structure of the icy crust. Thermal conduction is the main heat transport in regions where no other transport mechanism is operating (i.e. in cold lithospheres). It plays also a major role in regions where thermal gradients are large, for instance in the vicinity of the thermal anomalies associated with the tiger stripes (Abramov and Spencer 2009). However, the thermal conductivity of water ice is not large enough to efficiently remove the heat generated within Enceladus, so that other more efficient transport mechanisms are required.

The intense plume activity on Enceladus suggest that vapor production and circulation through the ice is an efficient mechanism to transport heat to the surface (Nimmo et al. 2007) (Fig. 3). However, such a transport mechanism can efficiently operate only in regions with high porosity, therefore mainly in the upper crust where the confining pressure is low. It is conceivable that vapor can be produced at shallow depths (down to few hundred meters), but it gets rather unlikely at deeper levels. Indeed, water vapor can be produced from sublimation
of water ice only if the vapor pressure remains at or below the saturation (or equilibrium) vapor pressure. For $T = 200$ K, the saturating vapor pressure is only 0.16 Pa (e.g. Schorghofer [2007]). As a comparison, the hydrostatic pressure at one kilometer below the surface is of the order of 100 kPa. Maintaining vapor pressures as low as 0.2 Pa within pores a few hundred meters and more below the surface would require all pores to be highly interconnected, resulting in an extremely fractured permeable medium. Pressures in active wide fractures could possibly remain close to the near-vacuum surface conditions under the action of tensile tidal stresses if vapor is instantaneously removed (Kieffer et al 2006; Hurford et al 2007). However, this is possible only locally in the first hundred meters below the surface. Therefore, vapor production and transport can be considered as a major factor in advection of heat only near the surface.

Below the porous and fractured upper crust, heat is more likely advected through thermal convective instabilities of solid ice and possibly by circulation of water coming from the underlying ocean. Solid-state convection in Enceladus' ice shell is predicted to occur only if the viscosity of ice at the base of the ice shell is less than $10^{14}$ Pa s, corresponding to rather small grain size ($< 0.3$ mm, Barr and McKinnon 2007) if the deformation of ice is controlled by grain boundary sliding. The efficiency of heat removal by solid-state convection depend not only on the basal viscosity but also on the mobility of the upper layer (Barr 2008; Roberts and Nimmo 2008; Tobie et al 2008; O’Neill and Nimmo 2010; Běhounková et al 2010) (Fig. 4c). In the asymptotic stagnant lid regime, scaling laws for a volumetrically heated fluid (Grasset and Parmentier 1998; Solomatov and Moresi 2000) predict that the heat flux ranging between roughly 15 and 30 mW m$^{-2}$. This can explain less than 20% of the heat flux observed at the SPT. Moreover, convection in the...
standard stagnant lid regime is unable to reproduce the lateral variations as observed in the thermal emission data (Spencer et al. 2006; Abramov and Spencer 2009). Such a convection regime does not take into account the mobility of the overlying lithosphere, which is likely to participate in the convective cycle (Barr 2008; Tobie et al. 2008; O’Neill and Nimmo 2010). On Earth, for instance, the motions of lithospheric plates play a crucial role in the advection of heat (e.g., Sotin and Parmentier 1989). Plate spreading at terrestrial midocean ridges is known to influence the upwelling of warm mantle materials and hence increase the release of internal heat (Sotin and Parmentier 1989). Although the tectonic processes on Enceladus are probably different, active spreading can also be envisaged over Enceladus’ south polar region (McKinnon and Barr 2007). An average spreading rate of several centimeters would be sufficient to explain the observed thermal emission at the south pole (McKinnon and Barr 2007). The spreading would also help focus the streamlines below the ridges (Sotin and Parmentier 1989), leading to a convergence of warm materials heated over a relatively broad dissipative region toward the narrower tectonic active region at the south pole.

An alternative heat transport may be closed-loop circulation of water from a subsurface ocean (Matson et al. 2010) (Fig. 4b). Ocean water is denser than the icy crust and normally cannot rise to the surface. Liquid water can, however, be injected at the base of the ice layer under the action of tidal forces leading to the formation of bottom crevasses (Mitri et al. 2008) and owing to overpressurization of a crystallizing internal ocean (Manga and Wang 2007). As water moves upward owing to overpressurization, the pressure is reduced and gases initially dissolved in water can exsolve, form bubbles, and make the ocean water less dense than the surrounding ice (Crawford and Stevenson 1988). Buoyant ocean water could therefore rise toward the surface and carry with it heat and chemicals that are delivered to the plume chambers and the shallow reservoirs that keep the surface ice warm. Depleted of heat and gas bubbles, the water becomes more dense and descends through cracks that return it to the ocean. Once the buoyancy has been sufficiently established the closed-loop circulation is self-sustaining. The heat delivery rate is controlled by the rate at which water arrives and by how much heat it carries. A discharge of ~2000 m${}^3$s$^{-1}$ of seawater would be required to deliver the observed 15 GW (Howett et al. 2011). This can be compared with the supply needs of the plumes. The amount of mass ejected by the plumes has been estimated to be about 150 to 350 kg s$^{-1}$ whereas the heat flow needs a much larger mass of water, namely about a hundred tons per second. Another way of gauging the reasonableness of the supply requirement is to calculate how much of the surface area must be allocated to the upwelling of ocean water. Assuming a flow velocity of a centimeter per second, the actual area is relatively small, being only several millionths of the area of the South Polar Region (that has a radius of about 50 km). Higher flow velocities require less area.

Except for thermal conduction, the three other heat transport mechanisms require very specific conditions to operate. Solid-state convection in the ice shell can be initiated only if the viscosity is lower than 10$^{14}$ Pa.s, which is possible, but require small grain sizes ($<1$ mm) and temperature near the melting in most of the convecting zone. Maintaining small grain sizes in a convective ice shell is problematic as normal grain growth is relatively rapid at elevated temperature (Montagnat and Duval 2000; Barr and McKinnon 2007). Circulation of water is also problematic because as mentioned above liquid water is naturally denser than water ice. The injection of liquid water at the bottom of the ice shell requires relatively large tidal stresses or rapid ocean crystallization, and the extraction of liquid water up to the surface, or at least at shallow depths, therefore requires the presence of dissolved gases in sufficient amounts (at least a few 0.1 wt%, Tobie et al. 2010). Finally the vapor transport model can operate only in regions with very high porosity, therefore it is limited to the first hundred meters below the surface where the confining pressure is low. These different transport mechanisms are probably operating simultaneously, but it is unclear which of these transport mechanisms is the dominant one at present at Enceladus’ south pole, and how the efficiency of heat transport evolves as a function of time.
2.3 Evolution

2.3.1 Geological history

Enceladus is an active icy satellite operating geological processes that are also thought to have occurred on other icy satellites. Thus, Enceladus is important to better understand the histories of other icy satellites in the Solar System. Enceladus exhibits the greatest geological diversity in a complex stress and strain history that is predominantly tectonic and cryovolcanic in origin, with both processes stratigraphically correlated in space and time, although the energy source is so far not completely understood (e.g. Smith et al. 1982; Kargel and Pozio 1996; Porco et al. 2006; Nimmo and Pappalardo 2006; Nimmo et al. 2007; Jaumann et al. 2008, 2009; Spencer et al. 2009).

Three different terrain types exist on Enceladus. Ancient cratered terrains cover a broad band from 0° to 180° longitude. Centred on 90° and 270° longitude there are younger, deformed terrains roughly 90° wide at the equator. The southern polar region south of 55° is heavily deformed and almost uncratered. The most remarkable known aspect of Enceladus is its south polar plume activity (Hansen et al. 2006; Waite et al. 2006; Porco et al. 2006; Dougherty et al. 2006; Spahn et al. 2006b). This is the only unambiguous example of active cryovolcanism in the solar system. The plumes arise from warm surface fractures (Porco et al. 2006), the "tiger stripes" (Porco et al. 2006). Sublimation of warm surface ice (Spahn et al. 2006), boiling of near-surface liquid water (Porco et al. 2006), decomposition of clathrates at depth (Kieffer et al. 2006), and sublimation at depth due to frictional heating (Nimmo et al. 2007) have all been proposed as plume generation mechanisms.

The surface fractures radiate ∼15 GW (Howett et al. 2011) and the plume latent heat carries away ∼1 GW, and this energy must be continually resupplied by some heat source. Characterization of the plume is crucial to understanding the mechanism that maintains their activity. Currently ice particle production and escape rates are not well constrained because of limited knowledge of the plume particle size distribution (Porco et al. 2006; Spahn et al. 2006b). However, the mass production rate of the plume gas has been estimated to be about 120-180 kg/s from occultation data (Tian et al. 2007). This value is surprisingly high, sufficient to remove a significant fraction (∼20%) of Enceladus’ mass over the age of the solar system (Kargel 2006). These tiger stripes are typically ∼500m deep, ∼2 km wide, up to ∼130 km long and spaced ∼35 km apart. Crosscutting fractures and ridges with almost no superimposed impact craters characterize the area between the sulci. The tiger stripes are apparently the source of the geysers observed to emanate from the south polar region (Spitale and Porco 2007; Spencer et al. 2009). Tectonic features dominate the surface, demonstrating, that also smooth terrain is pervasively tectonized (see section 2.1.4). Several different tectonic processes seem to have been at work. The sinuous chain of scarps that bound the south polar terrain at a latitude of ∼55° appear to have formed in response to compressional forces, while north-trending fracture zones that radiate from peculiar Y-shaped cusps and interrupt the chain of scarps appear to be extensional (Porco et al. 2006). Shear offsets along pre-existing rifts are also observed near the transition between these contractional and extensional features.

While the south polar terrain is a focus of pervasive active tectonism, other regions are not (e.g., the cratered north polar region). Analysis of the relationship between impact craters and tectonic features (Barnash et al. 2006; Bray et al. 2007) indicates that the tectonism has persisted through time. Furthermore, fossil terrains elsewhere on Enceladus reminiscent of the south polar terrain suggest multiple resurfacing episodes throughout the satellite’s history (see section 2.1.4).

The chain of scarps bounding the south polar terrain, as well as the northward-radiating fracture zones, may be the product of a change in Enceladus’s global figure. The coherent (but latitudinally asymmetrical) global pattern of deformation on Enceladus is currently unexplained, but it is most probably due to shape
changes, perhaps related to a hypothetical episode of true polar wander (Nimmo and Pappalardo 2006).

2.3.2 Internal dynamics and reorientation

Several observational constraints suggest that the ice shell of Enceladus is dynamically active, at least in the southern hemisphere. The existence of the young terrain (tiger stripes) at the South Pole may be evidence for mobile lid tectonics, perhaps associated with convection (Barr 2008). Similarly, the high polar heat flux may be due to episodic overturn of a convecting ice shell (O’Neill and Nimmo 2010). Moreover, the localization of heat production by tidal friction at the south pole may be attributed to a regional thermal anomaly above a liquid water reservoir (Tobie et al 2008). Outside the south polar terrain, several theoretical studies showed that thermal convection is possible (Barr and McKinnon 2007; Roberts and Nimmo 2008; Behounková et al 2010). However, no observation permits us to confirm the occurrence of convection on a global scale.

Stereo topography shows that the surface of Enceladus is notably bumpy at the ∼100 km lengthscale (Schenk and McKinnon 2009) (see section 2.1.1). The ∼1 km amplitude of some of the depressions is too large to be plausibly due to thermal convection, even in the event of ice shell yielding (Showman and Han 2005). On the other hand, shell density or thickness variations could easily explain such amplitudes. Shell thickness variations would be rapidly erased if convection takes place within the shell, but could persist for longer periods in a conductive shell (Collins and Goodman 2007). Density variations in the cold, near-surface lid could potentially co-exist with deeper convection, but would need to be large (δρ/ρ ∼ 0.1) to explain the topography anomalies. Perhaps the simplest explanation for the observed topography is that the ice shell of Enceladus is static and conductive, except near the south pole.

The polar location of the hot spot is plausibly due to reorientation of the satellite driven either by a low-density blob (diapir) within the silicate core or the ice shell (Nimmo and Pappalardo 2000), or by the topographic low at the south pole (Collins and Goodman 2007). In the diapir case, motion is not necessary; all that is required is a density contrast. Ice shell density contrasts needed for reorientation are large, more suggestive of compositional variations than thermal variations. However, compositional variations could be induced by melting caused by thermal variations (Pappalardo and Barr 2004; Tobie et al 2008). Studies of coupled thermal-compositional convection on Enceladus are currently in their infancy (Stegman et al 2009). It has been suggested that the global tectonic pattern on Enceladus might be the result of a ∼90° reorientation episode (Matsuyama and Nimmo 2008). However, the spatial density distribution of crater does not seem consistent with recent polar wander (Kirchoff and Schenk 2009). More detailed geological mapping, summarized in Spencer et al (2009), lends some credence to this idea, although another mechanism (such as diurnal tides) has to be invoked to explain the apparently tensional nature of the tiger stripes. The best evidence that reorientation is likely to have occurred would be the detection of a negative gravity anomaly centred on the south pole (Nimmo and Pappalardo 2006).

2.3.3 Ocean lifetimes

The lifetime of any subsurface water depends on how rapidly heat is being extracted into the overlying shell, and how much heat is being added to the water either from internal dissipation or by the silicate core beneath. Because there is expected to be little heating in the core, and the shell of Enceladus is relatively thin, predicted global ocean lifetimes are only a few tens of Myr (Roberts and Nimmo 2008).

A simple analytical treatment illustrates the problem. In steady state, the temperature at the base of a spherically-symmetric, bottom-heated conductive shell is given by:

\[ T_b - T_s = \frac{F_b}{k} (R - fR) \]  (1)
where $T_s$ is the surface temperature (80 K), $F_b$ is the heat flux at the base of the shell, $k$ is the thermal conductivity (assumed constant), and $R$ and $fR$ are the outer and inner shell radii, respectively.

If tidal heating is concentrated in the warmest, basal ice (or in the ocean below; Tyler (2011)) then equation (1) will provide a reasonable estimate of the basal ice temperature (assuming spherical symmetry). Heating distributed throughout the shell would result in a lower temperature, and convection will only increase the rate of heat loss, so this approach is conservative. The steady state tidal heating contribution is about 1.1 GW (assuming $Q = 18000$ in Saturn, Meyer and Wisdom (2007)), resulting in a basal heat flux of about $4 \text{ mW m}^{-2}$ for $f = 0.6$; to this must be added the silicate contribution of about $1 \text{ mW m}^{-2}$ (Schubert et al 2007).

Application of equation (1) and taking $k=3 \text{ W m}^{-1} \text{K}^{-1}$ then results in a basal temperature of about 180 K. This result shows that steady-state tidal heating irrespective of where it is being generated – is too small to permit a global ocean to survive, except in cases of extreme melting point depression (the ammonia-water eutectic freezes at 177 K).

On the other hand, a regional sea can be sustained. Although equation (1) will only be approximately correct if the heating varies spatially, it does suggest that increasing the local basal heat flux by a factor $\sim 2$ would be sufficient to maintain a local water body in steady state. Spatial variations in tidal heating are likely at Enceladus (e.g. a warm blob at the south pole will focus dissipation), and detailed numerical modelling indicates that a regional sea can indeed persist in quasi-steady state (Tobie et al 2008; Behounková et al 2010).

In summary, a long-lived regional body of water seems more likely than a short-lived global ocean. If complete freezing ever takes place, dissipation in Enceladus as a whole will decrease significantly, and equation (1) shows that it is then very difficult to re-melt the ice.

2.3.4 Coupled thermal-orbital evolution and episodic activities

A critical constraint on the evolution of Enceladus is that the observed current rate of heat loss significantly exceeds the rate (1.1 GW, assuming $Q = 18000$ in Saturn) which can be maintained in steady state by tidal dissipation. Other present-day sources of heat (e.g. radioactivity) are small, and primordial sources (e.g. accretion, $^{26}\text{Al}$ decay) will have been lost long ago.

There are thus only two possibilities: Enceladus is generating heat at the steady-state rate, but the rate of heat loss varies with time; or heat production itself varies with time. Either eventuality means that we are seeing Enceladus at a special time.

Episodic heat loss can arise as a natural consequence of convection, associated to episodic resurfacing events (Tobie et al 2008, O’Neill and Nimmo 2010). Fig 5a shows the surface heat flux for a convecting fluid with a constant internal heating rate and a relatively weak lid, as predicted by O’Neill and Nimmo (2010). Every so often, the convective stresses become large enough to cause localized yielding and downwelling of the lid, bringing warm ice to the surface and resulting in a pulse of heat. The heat pulses last $\sim 10 \text{ Myr}$ and are separated by intervals of quiescence of several hundred Myr. In this case, the heat production rate is the steady-state value and thus Enceladus’ eccentricity need not have changed at all. This scenario, although undoubtedly not correct in detail, thus provides one way of explaining Enceladus’ evolution.

Alternatively, Enceladus may have undergone oscillations due to the coupling between eccentricity, tidal heating and thermal evolution (cf. Hussmann and Spohn 2004). However, the eccentricity damping timescale is fast compared to the likely timescale for heat transfer across Enceladus. Thus, the kind of oscillatory behaviour proposed by Ojakangas and Stevenson (1986) for Io does not occur at Enceladus (Meyer and Wisdom 2008), irrespective of whether the heat transfer is convective or conductive. Other possibilities do exist, however. For instance, Fig 5b shows a scenario similar to that proposed by Stevenson (2008). Here Enceladus is initially in a non-dissipative state, so the eccentricity increases; once the resulting
diurnal tidal stresses exceed some critical value, fault motion and shear heating initiate, and dissipation then drives the eccentricity back down. Once a different threshold value is reached, fault motion ceases and the cycle repeats. In this case, both the eccentricity and the heat production vary with time. Similar to Fig. 5a, the heat pulses are short compared to the intervals of quiescence.

How may we distinguish between these two possibilities? One obvious difference is that Fig. 5b requires the eccentricity to vary with time. The rate of eccentricity damping at Enceladus is given by (e.g., Sohl et al. 1995):

$$\frac{de}{dt} = -\frac{a(1-e^2)}{GM_SM_T} P_{tide},$$

where $e$ is eccentricity, $a$ is semi-major axis, $P_{tide}$ is the tidal dissipation rate and $M_S$ and $M_T$ are the masses of Saturn and Enceladus. For a present-day dissipation rate of 15 GW, the eccentricity damping timescale is about 0.8 Myr and the timescale for the change in mean motion is about $7.5 \times 10^{-11}$ yr$^{-1}$. Such rates are potentially detectable, given a sufficiently long baseline of observations as it was done for Io (Lainey et al. 2009). However, the change in eccentricity and mean motion depends also on the dissipation inside Saturn, which counterbalance the effect of Enceladus’ dissipation. Preliminary analyses by Lainey et al. (2010) indicates that tidal dissipation in Saturn is about ten times larger than what was initially anticipated. If this is confirmed, it would indicate that Enceladus may dissipate up to 10-12 GW without damping its eccentricity. This would imply that Enceladus is close to equilibrium tidal heating, and that episodic activity is no longer needed.

Clearly, understanding the long-term evolution of Enceladus is a difficult problem, involving a non-spherically-symmetric body in which the thermal and orbital behaviour are tightly coupled. So far relatively little work has been done on this topic. One exception is Zhang and Nimmo (2009), which examined the orbital evolution of Enceladus and Dione and concluded that a convective Enceladus would be too dissipative to permit the present-day characteristics of the Enceladus:Dione resonance. However, all the previous works on orbital evolution would need to be revised if high dissipation inside Saturn is confirmed. This is an area in which future work is likely to prove fruitful.

3 Geochemistry

3.1 Chemical observations

3.1.1 Surface composition

The surface of Enceladus is mainly composed of water ice. This is evidenced from observations at the visible and near infrared (0.8-5.1 µm). Cruikshank (1980); Bu-
ratti and Veverka (1984); Cruikshank et al (2005); Emery et al (2005); Brown et al (2006); Verbiscer et al (2006); Newman et al (2008); Filacchione et al (2010), and far ultraviolet (115-190 nm, Hendrix et al (2010)) wavelengths. The exceptionally high geometric albedo of the moon (Buratti and Veverka 1984; Verbiscer et al 2005) is consistent with these interpretations. The dominance of water ice grains in the plume’s solid emissions (Porco et al 2006; Hedman et al 2009; Teolis et al 2010; Postberg et al 2011) and the E ring (Hillier et al 2007; Postberg et al 2008; 2009) implies that the moon’s surface is partially covered by a thin layer of icy grains emitted from the subsurface. The deposition of icy grains on Enceladus surface is observed and described (Kempf et al 2010; Schenk et al 2011; Degruyter and Manga 2011). Disc-integrated spectra of the moon indicate the dominance of crystalline water ice (Grundy et al 1999; Emery et al 2005). Photometric and spectral analysis of Cassini near infrared reflectance spectra also reveals the dominance of crystalline ice except the areas between linear tectonic depressions (the tiger stripes) at the south polar region (Brown et al 2006; Newman et al 2008). These local areas demonstrate the largest signature of amorphous ice. However, a high degree of ice crystallinity is observed at the warm tiger stripes and could be caused by crystallization of amorphous ice (Newman et al 2008).

Other surface compounds are not abundant and mainly imbedded in water ice. Traces of CO$_2$ complexed with ice and light organic compounds at the tiger stripes are reported based on Cassini near infrared observations (Brown et al 2006). Free CO$_2$ is also seen northward of the tiger stripes region (Brown et al 2006). Ammonia and/or ammonia hydrates have been suggested as possible trace (<1%) impurities based on interpretation of near infrared (Emery et al 2005; Verbiscer et al 2006) and ultraviolet spectra (Hendrix et al 2010). These interpretations are consistent with the detection of CO$_2$, NH$_3$, and organic species in plume gases (Waite et al 2006, 2009, 2011). The occurrence of high molecular weight organic molecules in surface materials agrees with the likely presence of those species in the plume and E ring particles (Postberg et al 2008). Correspondingly, the dominance of salt-bearing icy particles in solid plume emissions (Postberg et al 2011); Section 3.1.3) implies the presence of salts in the south polar surface layer composed of deposited plume particles. Re-accretion of the plume and E ring salt-bearing particles (Section 3.1.3) also suggests a global occurrence of Na-, K-, Cl-, and carbonate species in the surface materials.

3.1.2 Composition of plume gases

The composition of the plume gases has been measured in situ by the Ion and Neutral Mass Spectrometer (INMS) (Waite et al 2004, 2006, 2009) and remotely by the Ultraviolet Imaging System (UVIS) (Esposito et al 2004; Hansen et al 2006, 2008, 2011). In addition, particle measurements by the Cassini Dust Analyzer (CDA) in the E ring and in the plume (Srama et al 2004; Postberg et al 2009, 2011), the near-Enceladus plasma environment (Smith et al 2005, 2007, 2008; Christon et al 2013) and surface composition (Brown et al 2006; Hodyss et al 2009; Hendrix et al 2010) provide indirect evidence of the plume’s composition. The first measurements of plume composition were made with the INMS and UVIS instruments during the second targeted Enceladus flyby (E2 Tab.1) on 14 July 2005 and revealed that the plume is composed predominantly of water. The E2 flyby was an upstream pass, during which the Orbiter only grazed the plume’s low-density outer edge, with a minimum distance of 330 km to the plume sources. INMS pointing during the pass was not favorable to achieving the maximum signal-to-noise ratio. Despite these less than ideal conditions, however, INMS was able to identify, in addition to H$_2$O as the dominant species, significant amounts of CO$_2$ and CH$_4$ in the plume as well as an unidentified species with a mass-to-charge (m/z) ratio of 28 (Waite et al 2006), which was initially thought to be either CO or N$_2$. The INMS E2 data also indicated the possible presence of trace quantities of C$_2$H$_2$, C$_3$H$_8$, NH$_3$, and HCN. Mixing ratios relative to H$_2$O for the most securely identified species are: CO$_2$ 0.6% (+/- 0.15%), CH$_4$ 0.23% (+/- 0.06%), and...
UVIS occultations of Enceladus’ plume have yielded compositional results complementary to the INMS measurements. The first UVIS measurement of the plume came from the stellar occultation occurring on July 14, 2005 in conjunction with the E2 encounter. Absorbance detected in the far ultraviolet (FUV, 111.5-191.4 nm) channel confirmed H2O dominance in the plume gas (Hansen et al 2006). The FUV channel offered a good opportunity to observe the presence of CO and help clarify the INMS ambiguity at 28 Da. The measured absorption spectrum, however, showed no sign of CO’s known absorption bands, leading Hansen et al (2006) to set an upper limit mixing ratio of 1%. The second UVIS plume observation was obtained during a stellar occultation on October 24, 2007 where absorption spectra in the FUV channel again failed to show signs of CO (Hansen et al 2008). The non-detection of CO and formal 2σ upper limit mixing ratio of 3% relative to H2O set by Hansen et al (2008). CO was also not detected in the Visible and Infrared Mapping Spectrometer (VIMS) surface reflectance spectra from Enceladus’ south polar region (Brown et al 2006). The most recent UVIS observation of Enceladus plume comes from the solar occultation event on May 18, 2010. The extreme ultraviolet (EUV, 55-110 nm) channel was utilized for this observation, allowing for possible detection of N2. The UVIS absorption spectrum showed that that N2 absorption feature was limited to a level corresponding to the signal standard error, constraining the abundance of N2 to an upper limit of 0.5% (Hansen et al 2011). Absorption cross-sections are temperature sensitive, so quantitative abundance determinations of other species from UVIS data would require a full spectrum absorption cross section for H2O below 200 K.

Analysis of high-sensitivity INMS spectra from the high-speed E5 flyby, during which the Orbiter traveled closer to the plume center on a north-south trajectory following the flow direction of the plume (Tab.1), suggested that the m/z 28 signal was due to CO produced in the detector by the hypervelocity impact dissociation of CO2 plus a smaller amount of either N2 or C2H4; any contribution to the signal from endogenous CO was considered to be small and indistinguishable from the dissociation product (Waite et al 2009). However, analysis of data acquired during the low-altitude, low-speed E7 pass directly through the plume center (Tab.1) suggested that CO, if present, cannot be fully explained by CO2 impact dissociation and any CO native to the plume (Waite et al 2011). According to one of the compositional scenarios discussed by Waite et al (2009), if N2 is assumed to be present in the plume, then HCN must also be present. However, the detection of HCN reported by Waite et al (2011) based on the E7 data has not been confirmed by data from subsequent low-altitude flybys; within statistical uncertainties, the m/z 28 signal can be accounted for by several combinations of species (e.g., CO, N2, HCN, C2H4) (Waite et al 2009; Supplementary Information, and Magee et al., paper in preparation, 2013).

N+ detected by the Cassini Plasma Spectrometer (CAPS) in Saturn’s inner magnetosphere (Smith et al 2005) could in principle indicate the presence of N2 (as well as of NH3) in the plume (Smith et al 2007). There has been no definitive detection of the N2+ parent ion, however, and the CAPS data permit only an upper limit to be established for it (Smith et al 2008).

Suprathermal mass-28 ions have been detected with the CHarge-Energy-Mass Spectrometer (CHEMS), but the contributing species—whether N2+, CO+, or some other species—cannot be resolved (Christon et al 2013).

Based on the data from the E5 flyby, Waite et al (2009) reported the definitive detection of NH3 and the measurement of a deuterium-to-hydrogen ratio in the plume H2O close to the value determined for a number of comets (but see also Hartogh et al (2011)) and twice the terrestrial D/H ratio. The presence of C4 and heavier hydrocarbons and the tentative detection of 40Ar as well as of formaldehyde and methanol were also reported. However, INMS data from the E7 and later flybys during which the Orbiter passed at low altitudes directly through the center
of the plume provided no evidence for $^{36}$Ar and, moreover, showed substantially reduced abundances for the majority of the trace species reported by Waite et al. (2009) (Figure 6). Spectral signatures for $C_3$ hydrocarbons are clearly evident in the low-altitude data sets, although there are ambiguities in the identification of the individual $C_3$ hydrocarbons because different combinations of species can be assigned to a particular m/z value according to the compositional model selected to fit the data. Only upper limits can be given for heavier hydrocarbons (such as $C_6H_6$). The E7 and later low-altitude flybys occurred at velocities relative to Enceladus that were a little less than half that of the E5 flyby (Tab. 1). Waite et al. (2011) thus interpreted the compositional differences between the E5 and the low-speed data sets in terms of a velocity effect involving the impact dissociation of certain molecules at the higher kinetic energies of the E5 flyby. For example, the signatures of $C_4$ and heavier hydrocarbons seen in the E5 data but not in the E7 data may indicate the presence of fragments of larger, more complex organic molecules (with masses outside the INMS 100-Da range) that break up upon impact with the surface of the instrument’s antechamber.

Other lines of circumstantial evidence also point to the likely presence of organic macromolecules and simpler hydrocarbons in the plume. Mass-spectral analysis of E-ring particles believed to originate in Enceladus’ plume (Type II particles) indicates that they contain silicates and/or fragments of refractory organics such as “polar organic compounds (e.g., alcohols, nitriles, acids) or long-chain hydrocarbons” (Postberg et al. 2008; see also section 3.1.3). Additional evidence comes from surface reflectance data at various wavelengths (cf. Section 3.1.1). Absorption features in the VIMS spectra at 3.44 $\mu m$ and 3.53 $\mu m$ have been interpreted as signatures of short-chain organics present in the tiger stripes (Brown et al. 2006). Hodyss et al. (2009) attributed the feature at 3.53 $\mu m$ to methanol, which is either produced photochemically on the surface of Enceladus or released in the plume and deposited on the surface (INMS data from the slow flybys allow only a conservative upper limit of < 0.004% to be established for methanol [Magee et al., paper in preparation, 2013]). Hendrix et al. (2010) found that the best fit to spectra of Enceladus’ surface at FUV, middle-ultraviolet, and visible wavelengths is achieved with spectral models that combine small amounts of NH$_3$ and tholin with H$_2$O. An H$_2$O-NH$_3$-tholin model spectrum was also recently used by Zastrow et al. (2012) to fit a composite Enceladus reflectance spectrum generated from Cassini UVIS and HST COS and STIS data. To account for the possible presence of tholins in Enceladus’ ice, Hendrix et al. (2010) suggested that the “complex hydrocarbons” reported by Waite et al. (2009) “could undergo processing to result in the tholin-type material predicted to be present.”

### 3.1.3 E ring and plume particles

The Saturnian system is swamped by small particles (size < 10 $\mu m$) escaping from Enceladus gravitational domain, most of them confined within Saturn’s outermost diffuse E ring. The Cosmic Dust Analyser (CDA) onboard the Cassini spacecraft allows for chemical characterization of these particles. The instrument produces time-of-flight mass spectra from cations and cationic aggregates of the gas and plasma cloud generated by high-velocity impacts of single grains onto a metal target (Srama et al. 2004). CDA carried out compositional measurements of both E ring particles (Hillier et al. 2007; Postberg et al. 2008, 2009) and freshly ejected particles during plume crossings (Postberg et al. 2011). Another dust family of Enceladus origin are the Saturnian stream particles (Kempf et al. 2005, 2006; Hsu et al. 2010, 2011a,b). These are tiny nanometer-sized grains whose dynamics are not governed by gravity but by electromagnetic interaction. Spectra recorded in the E ring are best for compositional analysis, as the spacecraft spent much more time recording spectra while crossing the E ring than in the plume. Thus, E ring spectra have an excellent statistical basis. Moreover, the dust density during plume crossings resulted in such extreme impact rates that the stress on the instrument led to disturbances of the spectral signal, lowering the compositional information of individual grains.
Enceladus as an active body

Fig. 6 The accumulated INMS signal mass spectra for E5 and E7 encounters of the Enceladus plume representative of the high and low velocity distributions. The spectra have been normalized so that the signal in other masses can be compared to the H$_2$O signal at 18 Da. The lower bound of the plot window represents the noise level for both data sets. A significant dependence on spacecraft velocity is observed when comparing the two spectra. The E5 spectrum shows composition with complexity up to C$_6$H$_6$ (78 Da) whereas E7 complexity peaks in the C$_3$ group (around 40 Da). (Note: These spectra are presented merely to illustrate the velocity effect observed in the INMS data and should not be used for detailed analysis. New spectra and mixing ratios based on an extensive re-analysis of the INMS plume data from all of the relevant Enceladus flybys are currently in preparation.)

E ring grains Even before Cassini’s arrival it was suspected that water plays an important role in the E ring’s composition. However, only Cassini revealed that Saturn’s environment outside the main rings is totally governed by water ice particles (Hillier et al 2007) as well as water molecules and ions which form from water dissociation (Young et al 2005; Esposito et al 2005; Melin et al 2009). Almost all particles encountered in the E ring are dominated by water ice but nevertheless there are significant differences in their individual composition. E-ring impact spectra obtained with CDA can be categorized into at least three distinct families (Postberg et al 2008, 2009). Type I spectra imply almost pure water ice grains whereas Type II spectra exhibit contributions from organic and/or siliceous material. Type III spectra show a drastically increased alkali salt content. Although there are a few cases where individual spectra show characteristics of two types, they form distinct families.

Type I particles About 65% of all particles above the detection threshold belong to Type I. The abundance of this type increases further for small particles sizes. Mass lines caused by water cations (H$_2$O)$_n$(H$_3$O)$^+$, n = 0...15 are dominant or exclusive. Na$^+$ and K$^+$ and their respective water cluster ions (H$_2$O)$_n$(Na, K)$^+$ are present in most cases and form the only non-water mass lines. Calibration experiments point at very low concentrations of alkali salts with an Na/H$_2$O ratio in the order of 10$^{-7}$ (Postberg et al 2009).

Type II particles Type II spectra represent the second most abundant E ring family and show the same characteristic as Type I with an additional distinct feature between mass 27 and 31 which in most cases represents more than one mass line (Postberg et al 2008). In many cases additional non-water signatures appear and the peaks are broader than in the case of Type I or III. The additional peaks strongly point at unspecified organic compounds with strongly varying concentration. A contribution of silicates is also possible. However, water ice is always the dominant species.

Type III particles This family of about 6% of E ring spectra exhibit a totally different pattern of mass lines. In contrast to Type I and II, the water cluster peaks (H$_2$O)$_n$(H$_3$O)$^+$ are absent or barely recognisable. The characterizing mass lines are of the form (NaOH)$_n$(Na)$^+$, n = 0...4 indicating a Na/H$_2$O
Fig. 7 Composition and size profile of the Enceladus’ ice particle plume (modified after Postberg et al. [2011]). The background colours on the left panel show the fraction of salt-rich grains. Overlaid are contours of constant mean particle radius obtained from the model. The projection used is in the plane of the spacecraft trajectory during Cassini’s E5 flyby (solid black line, shown with 10 s intervals). CDA measurements show that both the largest particles and the highest fraction of salt-rich grains a few seconds after closest approach to Enceladus. Structures of the three most relevant localized supersonic jets for this projection are clearly visible in both the compositional profile and size contours. Note that the model only considers particle sizes above the instrument’s detection threshold (0.2 m) and it shows the plume emission with an added E ring background that is increasingly relevant north of the south polar source region. The right panel shows the modeled contribution from the few collimated supersonic jets versus more numerous slow, diffuse sources distributed all along the tiger stripes. Most importantly it shows the fraction of salt-rich ice as a function of logarithmic altitude above the south pole. Near the surface salt-rich particles account for about 70% of all grains > 0.2 \mu m but due to their larger average size they account for almost all the solid mass created by Enceladus’ active region.

Plume particles In 2008 and 2009 CDA carried out compositional analyses on three plume crossings, E4, E5, and E7. The majority of spectra can be assigned to one of the three types previously detected in the E ring. Whilst the number of obtained plume spectra was limited to fewer than twenty on E4 and E7, a specific configuration of CDA at E5 provided a spatial resolution high enough to infer a compositional profile of the plume (Fig. 7).

The amount of salt rich Type III grains increases above 40% at the closest approach to Enceladus’ surface (21 km). The observed compositional profile can be reproduced when Type III grains are ejected with slower speeds corresponding to their larger size. The best match points at about twice the average size of Type I grains (Postberg et al. [2011]). Applying a size dependent stratification of the plume (Schmidt et al. [2008]), which is in agreement with Cassini observations (Kempf et al. 2008; Hedman et al. 2009), about 70% of all ejected particles are of Type III. Owing to their larger size about 90% of the emitted solid mass must be salt rich ice (Fig. 7). Naturally the small and fast salt-poor particles preferentially escape the gravitational domain of Enceladus and therefore dominate the ice-population of the E ring.

mole ratio above $10^{-3}$. Frequent mass lines of NaCl – Na$^+$ and Na$_2$CO$_3$ – Na$^+$ reveal NaCl and NaHCO$_3$$^-$ and/or Na$_2$CO$_3$$^+$ as the main sodium bearing compounds. Calibration experiments indicate an average concentration of 0.5 – 2% sodium and potassium salts by mass with K compounds being less abundant by a factor of at least 100 (Postberg et al. 2009). The composition has a very good match to the composition predicted for liquid water on early Enceladus (Zolotov 2007) and implies a basic pH of 8 – 9. Impacts of Type III particles have an average ion yield which is several times higher than of Type I, implying a considerably larger size.
The fit could be further improved by assuming sources with drastically varying gas speeds. Hundreds of slower sources located all along the tiger stripes with gas speeds of about 400 m s\(^{-1}\) would preferentially produce salt rich grains. In contrast, the few collimated supersonic jets (Spitale and Porco 2007) observed in the UV and the optical wavelengths with gas speeds above 1000 m s\(^{-1}\) (Hansen et al 2008, 2011) are enriched with the smaller, sodium-poor particles (Postberg et al 2011). Furthermore, the CDA plume measurements imply a general enrichment of Type II ice grains containing organic material in the plume, probably associated with the fast collimated jets. Grains which are both enriched in salts and organic species are frequent.

**Saturnian stream particles** Cassini encountered these tiny (2 – 8 nm in radius), high-speed (> 100 km s\(^{-1}\)) particles in 2004 even before its orbit insertion at Saturn (Kempf et al 2005). The initial analysis of CDA impact spectra indicated little, if any, water ice and a major siliceous component (Kempf et al 2005). More surprising is the recent discovery that most stream particles start their dynamical life in the water dominated E ring (Hsu et al 2011a,b). This points at a stream particle release as refractory remnants of sputtered icy E ring grains in which they were initially embedded (Hsu et al 2011b). A thorough revision of stream particle spectra identified silica (SiO\(_2\)) as the main component of most stream particles. The most plausible explanation for this finding is the abundant formation of nano-colloidal silica in an aqueous phase which is then partly integrated in ice grains expelled by Enceladus’ plume (Hsu et al. 2013, in preparation). Small amounts of silica still embedded in ice grains with a Si/H\(_2\)O ratio below 10\(^{-4}\) would probably go undetected by CDA. However, occasional mass lines at 28 Dalton in spectra of E-ring and plume particles could be an indication for embedded nano-silica grains.

### 3.2 Composition of the interior

#### 3.2.1 Bulk composition and a rocky core

The density of Enceladus of 1608 ± 5 kg m\(^{-3}\) (Porco et al 2006) implies the presence of a rocky component in the interior. For the assumed densities of ice with some contaminants (1010 kg m\(^{-3}\)) and the rock (2500-3528 kg m\(^{-3}\)), the rock mass fraction is 0.62 to 0.52, respectively (Schubert et al 2007). In these evaluations, a rock density of 2500 kg m\(^{-3}\) roughly represents either the grain density of highly aqueously altered CI carbonaceous chondrites (Consolmagno et al 2008) or hydrated silicates, and the value of 3528 kg m\(^{-3}\) corresponds to the density of Io. A low rock density is likely because aqueous alteration is supported by the detection of salt-bearing grains emitted from the moon (section 3.1.3).

The rocky (inorganic) part of the moon could have the composition of CI carbonaceous chondrites very similar to that of the solar photosphere. However, the detection of typical cometary volatiles (CO\(_2\), NH\(_3\), and CH\(_4\)) in the plume, the comet-like abundances of NH\(_3\) and CH\(_4\), and D/H ratio in plume gases imply an enrichment of Enceladus in volatile and organic species compared to chondritic materials. Comets are strongly enriched in light elements relative to CI carbonaceous chondrites and the C-H-O-N type organic materials account for ∼23 wt% of the comet Halley dust (Jessberger et al 1988; Hanner and Bradley 2004). The likely presence of high molecular weight organic compounds in the plume and E-ring particles (sections 3.1.3 and 3.1.2) also agrees with an organic-rich interior of the moon. Thus the composition of Enceladus’ non-icy solids could be closer to the solar photosphere than to CI carbonaceous chondrites, which are depleted in C, O, and N. The density of hydrated organic rich rocks with the solar abundance of C could be ∼ 1900 ± 100 kg m\(^{-3}\) (Zolotov 2009).

In addition to abundant organic matter, the bulk density of rocks could be further lowered because of a high porosity. CI and CM carbonaceous chondrites have a porosity of 35% and 23%, respectively (Consolmagno et al 2008), and the pressure in the moon’s interior (<< 0.5 kbar) may not be sufficient to compact these
The thickness of the icy shell as a function of core density for a completely differentiated Enceladus (modified after Schubert et al. 2007). The error bars correspond to core density of 2500-4000 kg m\(^{-3}\) for hydrated and anhydrous rocks, respectively. An organic rich, porous, and hydrated core could have density below 2500 kg m\(^{-3}\). Differentiated Enceladus with such a low-density core could have a thin icy shell, as illustrated by arrows.

rocks below 10% porosity (Zolotov 2009). By analogy with chondrites (Consolmagno et al. 2008) this porosity could be mostly accounted for by microporosity so that the pores may not be filled with water and/or ices. If this is the case, it follows that a hydrated organic-rich rock with the solar abundance of C and 10% porosity could have a density of only \(\sim 1700\) kg m\(^{-3}\). This low end-member density corresponds to the rock mass fraction of 0.91. In a completely differentiated interior such a low density of the rocky core would correspond to a fairly thin icy shell (Fig. 8). Depending on the core density, the ice shell may be as thin as \(\sim 13\) km.

The low upper limit for \(N_2\) in plume gases (Hansen et al. 2011), a preservation of typical cometary species, and the low K/Na ratio in salt-bearing icy grains (Postberg et al. 2009) indicate predominantly low-temperature (\(T \leq 273\) K) water-rock interactions in Enceladus history (section 3.1). This implies a low amount of accreted \(^{26}\)Al and a formation of Enceladus at least several Ma after the Ca-Al-rich inclusions (CAIs), in agreement with recent formation models for the Saturnian system (Sasaki et al. 2010). The low-temperature rock alteration is in agreement with the global shape of Enceladus which is not consistent with hydrostatic equilibrium (Porco et al. 2006; Thomas et al. 2007). The shape suggests that the rocky core has never reached temperature high enough to relax core topography. In this condition, a large porosity may have persisted in the organic-rich and hydrated core, as discussed above.

Although some minor silicates may be present in icy particles emitted from Enceladus (Postberg et al. 2008), the data obtained with CDA are not sufficient to evaluate composition of Enceladus’ rocks. However, the supposed low-temperature alteration of the rocks implies that CM2 and some other carbonaceous chondrites (e.g., Zolensky and McSween 1988; Brearley and Jones 1998; Brearley 2006) could be used as a proxy for an altered mineralogy of the core. Inorganic solids could consist of abundant phyllosilicates (serpentine, cronstedtite, saponite, chlorite), primary and secondary sulfides (troilite, tochilinite, pyrrhotite, pentlandite), magnetite, phosphates, and carbonates. Many of these minerals are modeled to be present in an aequously altered Enceladus’ core (Zolotov 2007). The organic fraction of the rock could consist of aequously altered cometary-type C-H-O-N compounds (Hanner and Bradley 2004) and may also contain high molecular weight polymers abundant in carbonaceous chondrites (Septon 2002). Water-soluble and light organic compounds as well as high-solubility Na-rich salts could concentrate at the ice-core boundary along with \(NH_3\) and methanol that strongly decrease the melting temperature of ice (sections 3.2.3 and 3.1). \(CO_2\) from melted ices converts
Enceladus as an active body 27

Fig. 9 Stability curves for clathrate hydrates detected or suspected at Enceladus. The dashed curve corresponds to the mixed CO\textsubscript{2}-CH\textsubscript{4} clathrate composition at the CO\textsubscript{2}/CH\textsubscript{4} = 4. The red curves show modeled temperature profiles in the thermally conducting icy shell for different heat flow values (5 to 200 mW m\textsuperscript{-2}). The curve labeled 200\textsuperscript{*} corresponds to the surface temperatures of 200 K. Other red curves are for surface temperatures of 73 K. The red domain indicates the pressure-temperature conditions in a putative convecting ice layer. A temperature of 240 K is the typical value expected in cold downwelling, while hot upwellings typically have temperature of the order of 260-270 K. The blue area shows the stability field of liquid water.

The scheme in the lower part of the plot explains the conditions under which the clathrates are stable. The clathrate stability curves have been obtained with the CSMHYD software (Sloan and Koh 1998), except for the curve of CO\textsubscript{2}, which is determined from the August equation provided by Hersant et al (2004) for T < 273.15 K and from experimental data for T > 273.15 K (Mohammadi et al 2009).

3.2.2 Icy shell

As already discussed in sections 2.1.3 and 3.1.1, the surface of Enceladus is mostly composed of water ice, both crystalline and amorphous (Brown et al 2006). The plume is also dominated by H\textsubscript{2}O, in the form of vapor and ice particles (Hansen et al 2008; Waite et al 2009; Postberg et al 2011). However, the observation of non-water gas molecules in the gas plume (Waite et al 2009, 2011 section 3.1.2) as well as the detection of salt-bearing grains emanating from the south pole region (Postberg et al 2009, 2011 section 3.1.3) indicate the presence of impurities in the icy shell.

The gas molecules may be dissolved in liquid water (in a subsurface reservoir and/or localized melt pockets, e.g. Porco et al 2006), and/or trapped in the form of clathrate structure within the icy shell (e.g. Kieffer et al 2006). Clathrate hydrates are crystalline structures made of hydrogen-bonded H\textsubscript{2}O molecules, which form cages that can trap gas molecules (e.g. Sloan and Koh 1998; Lunine and Stevenson 1985; Choukroun et al 2013). The composition of common clathrates is G × n H\textsubscript{2}O, were G is a gas molecule and n ≈ 5.75. Clathrates occur in Earth’s permafrost, oceanic seafloor sediments and could store volatiles in various solar system bodies (see Choukroun et al 2013 for a review). On Enceladus, they have been proposed to play a dominant role in storing the gases, driving the plumes (Kieffer et al 2006) and controlling the tectonics by their stiff brittle nature (Gioia et al 2007a).
Concentrations of volatile species in the Enceladus plume compared to the Henry's law constants, which characterize solubility of gas species in pure water. The lack of clear correlation of plume concentrations with solubilities implies a significant contribution of plume gases from dry sources.

Fig. 9 illustrates the dissociation curves of several clathrate-forming compounds compared to typical temperature profiles expected in the icy shell, corresponding to different heat flow values. For all gas compounds, the dissociation temperature increases as a function of pressure (depth). H$_2$S clathrate is always stable even in the warmest temperature profile, corresponding to the conditions expected in the vicinity of the tiger stripes ($\Phi_S = 200$ mW m$^{-2}$, $T_S = 200$ K, where $\Phi_S$ and $T_S$ are the heat flow and temperature at the surface). The high stability of H$_2$S in the clathrate phase may explain its low abundance in the gas phase of the plumes. The other considered gas compounds are unstable in the form of clathrate for this warm case. For $200 > \Phi_S > 50$ mW m$^{-2}$, mixed CO$_2$ – CH$_4$ clathrates (having a CH$_4$/CO$_2$ ratio similar to the plume composition determined during the E7 flyby) become unstable at depth comprised between 3 and 10 km. For $T_S = 180$-200 K, mixed CO$_2$-CH$_4$ clathrate is always unstable. This implies destabilization of the CO$_2$-CH$_4$ clathrate in the vicinity of tiger stripes in the south polar region. Note that destabilization occurs even without decompression caused by fissuring in the ice. For $\Phi_S < 50$ mW m$^{-2}$, which may represent areas away from the south pole region, clathrates of pure CO$_2$ or CO$_2$-CH$_4$ mixture are always stable. Clathrates of CO and N$_2$ are stable only in very cold regions and are unexpected within the warm ice shell in the south polar region.

Despite the thermodynamic stability of clathrates, the role of clathrate dissociation as a main driver of plume activity remains a hypothesis. INMS data indicate that the major clathrate-forming gases (CO$_2$, CH$_4$) could compose less than $\sim$1 mole % of the H$_2$O dominated plume. If these gases are released through decomposition of clathrates plume source regions may contain less than 6 vol. % of clathrates. Such a low clathrate volume fraction suggests that clathrate crystals are mixed with ice crystals and probably do not form a spatially separated pure clathrate reservoir.

Gases which do not form stable clathrates in the icy shell (e.g. H$_2$) could be present in defects of ice crystals and in the spaces between the crystals. NH$_3$ and CH$_3$OH detected in the plume are very soluble in liquid water (Fig. 10) and do not form clathrates. These molecules, especially strong anti-freezing species NH$_3$ and CH$_3$OH, are likely to be concentrated in a potential liquid phase (section 3.2.3).

The degree of compositional heterogeneity of the icy shell is unknown. The diverse nature of chemical species emitted from Enceladus may indicate spatial separation of salt- and volatile-bearing regions of the shell. Pure water ice zones could
also be separated from salt-bearing water ices and volatile-bearing regions, which may represent newer melted primordial icy materials. The composition structure below the south polar region could be different from other regions. The compositional heterogeneity should reflect mechanisms of water-rock differentiation and subsequent thermal processes in the shell (sections 3.1.1 and 3.1.2). In one possible scenario, enriched liquids may be injected into the icy shell owing to the pressurization of a subsurface oceanic reservoir, resulting transient volume contraction of the liquid reservoir upon crystallization (Manga and Wang 2007). The dikes of enriched liquids may crystallize when reaching cold regions within the icy shell, leading to local enrichment in oceanic impurities (salts, NH$_3$, methanol, soluble and condensed organic species). Some oceanic materials could be concentrated in frozen lenses (cf., Matson et al. 2012). In another scenario, salts and organic compounds could be abundant in the upper horizons of the icy shell reflecting either ancient cryovolcanic activity or rapid freezing of early oceanic water after sinking of a primordial icy-rocky shell (section 3.1.1). In all these cases, the icy shell probably consists of a mixture of ice with local enrichment clathrate hydrate(s), organic compounds and salts concentrated upon crystallization of aqueous solutions.

### 3.2.3 Possible liquid phases

The stability of aqueous phases in the interior mainly depends on temperature and concentrations of species that decrease the melting temperature of ice. The energy output of the south polar region (>15 GW, section II.1.2) is sufficient to melt significant parts of the south polar ice. Although Cassini chemical data do not provide unequivocal evidence for large liquid water reservoirs, the occurrence of aqueous solutions in the interior is likely.

The observed surface temperature of ~190 K in the most active linear ridges of the south polar region (Spencer et al. 2011) is high enough to stabilize aqueous solutions even in surface materials. In fact, this temperature is higher than the temperatures of the ammonia-methanol-water (~153 K), methanol-water (~160 K), and ammonia-water (~175 K) eutectics (Kargel 1992). At the temperature of ~190 K, solutions rich in ammonia could exist as brine pockets in the surface ice. A high NH$_3$/CH$_3$OH ratio in plume gases implies a dominance of ammonia in brine pockets.

The subsurface temperatures at the Tiger Stripes must exceed the surface values (190 K) for any physically plausible scenario of heat-flow below the south polar terrain. Moreover, properties of the observed flow of vapor and ice particles (Schmidt et al. 2008) imply subsurface temperatures well above 240 K. These elevated temperatures would correspond to less concentrated and more abundant fluids in depth. The corresponding solutions could stably exist within ice as brine pockets rich in neutral species (ammonia, methanol, NaCl) and ions (Na$^+$, Cl$^-$, K$^+$, HCO$_3^-$, CO$_3^{2-}$, etc.) that both decrease the freezing temperature of an aqueous solution. The composition of fluid pockets should also reflect compositional heterogeneity in distribution of volatiles and salts of the icy shell. Depending on the origin of ices and temperature (sections 3.2.3, 3.3.1), the fluid pockets may be rich either in dissolved salts or in NH$_3$. If ammonia is not locally present, the near eutectic brines at the temperature ~238-243 K could be rich in Na$^+$ and HCO$_3^-$ while warmer fluids are NaCl dominated and less saline solutions (Fig. 4 in Zolotov 2007). If the composition of the emitted ice grains reflects the composition of the liquids they have an alkaline pH (Postberg et al. 2009), which depends on the temperature (Zolotov 2007).

The bottom of the icy shell is a place where current liquids are even more likely. The rock-ice interface could be enriched in species that decrease ice melting temperature. The enrichment could have occurred through downward freezing of a primordial ocean after differentiation (section 3.3.1). These impurities would contain species detected in plume gases and particles. In addition, organic compounds extracted from rocks could be abundant (section 3.3.1) and may separate in an organic liquid phase. The presence of these low-viscosity species would facilitate tidal motions and heating at the bottom of the icy shell (Zolotov 2007).
Tidal motions should favor ice melting and will decrease the salinity of aqueous fluids. In fact, a decoupling layer at the core-ice interface beneath the south polar region is required to generate enough tidal dissipation to account for the observed thermal emission \cite{Nimmo2007,Tobie2008}. Either a water reservoir or a low-viscosity organic-rich layer at the core-ice interface could serve as a lubricant that favors tidal flexing. In any case, the temperature at the ice interface needs to be close to the melting point of water ice to generate significant viscous dissipation in the overlying ice shell. A large liquid water reservoir (ocean) – if any, would be alkaline NaCl rich solution with subordinate amounts of CO$_3^{2-}$, HCO$_3^-$, and K$^+$ ions \cite{Zolotov2007,Postberg2009}. Lower volume of the reservoir would correspond to higher salinity. \cite{Glein2010} used mass balance calculations to evaluate NaCl content of this putative reservoir as a function of its geometry (volume).

Despite the possible existence of pore space in the rocky core (section 3.2.1), liquid phases may not exist within the rocky core. The depletion of long-lived radionuclides in rocks \cite{Schubert2007}, the lack of major tidal motions within rocks, and an accumulation of inorganic and organic impurities above the rocky core (section III.3.1) do not favor stability of aqueous fluids within the core. However, water-based and organic liquids could occupy pore spaces in the uppermost rock layer that underlie a liquid reservoir at the bottom of the icy shell.

### 3.2.4 Aqueous vs. dry sources of plume emissions

The striking compositional similarity of the E-ring salty ice grains \cite{Postberg2009} and solutions formed in water-rock interaction models \cite{Zolotov2007} indicates aqueous processes in Enceladus’ history. The dominance of salty ice grains in solid plume emissions \cite{Postberg2011} indicates a major contribution from aqueously processed materials. The question if the salty ice grains originate from ices or fluids requires a thorough consideration. This section discusses chemical species detected at Enceladus in the context of phase composition of plume sources.

The detected salt-rich plume grains \cite{Postberg2011} cannot originate from a salt-poor matter and must be emitted from salty ices or aqueous solutions. Some condensation of water on salty ice grains during their subsurface ascend implies higher concentrations of salts in the source region(s) than in the salt-rich grains forming the plume. The observed compositional uniformity of those grains (0.5-2 wt. % salt, \cite{Postberg2009}) requires a fairly homogenous composition of the source material, and plume sources should provide this homogeneity.

An aqueous source of salty ice grains is consistent with their compositional homogeneity. In one possible scenario, an aerosol spray forms through bursting of gas bubbles at the surface of subsurface aqueous reservoirs below back pressured vents \cite{Postberg2009}. The droplets freeze, thereby conserving homogeneous composition of the fluid, and are dragged upward by the gas flow. The salty grains formed from an aerosol spray would form larger ice grains than salt-poor grains which form by nucleation from water vapour \cite{Schmidt2008}. Therefore, this scenario is consistent with the observed larger size of salt-rich grains if compared with the salt-poor population \cite{Postberg2011}.

Evaporation of water from unconnected subsurface reservoirs would unevenly increase the fluid salt contents. However, even relatively small isolated reservoirs in the order of a few kilometer equivalent radius would need tens of thousands of years to change salt concentrations in the order of a few percent \cite{Postberg2009}. If these time scales are considered for the currently observed plume activity from subsurface, salty lakes’ are only consistent with large or connected water reservoir(s) below the south polar region.

It is harder to explain compositional homogeneity in terms of dry sources of salty grains. Sublimation of salty ices would lead to salt lag deposits at jet’s source regions and the sublimation could not account for the larger size of salt-rich grains. The earlier proposed origin of the plume’s icy grains through condensation after sublimation \cite{Nimmo2007,Kieffer2009} is not consistent
Enceladus as an active body

Concentrations of volatile species in the Enceladus plume (red symbols, see section 3.1.2) compared to the abundances in comets (blue boxes, data from Mumma and Charnley (2011), A’Hearn et al. (2012)). Note that CO$_2$ has not been reported in some comets. Plume concentrations of NH$_3$ and CH$_4$ closely match cometary data.

with the detection of salt-rich grains. Suggested downward freezing of an early salty ocean (see section 3.3.1) would have concentrated salts at the bottom of the forming icy shell and may not cause homogeneous trapping of salts in the main body of the icy shell. Even with flash freezing of early oceanic water it seems very difficult to form large source regions of compositionally homogeneous salty ice. Therefore, a comet-like violent sublimation which directly entrains ice particles in the flow may not produce compositionally homogeneous salty ice grains.

The proposed formation of icy grains through an explosive decomposition of clathrate hydrates (Kieffer et al. 2006; Gioia et al. 2007a) is not consistent with the small fraction of clathrate-forming plume gases (mainly CO$_2$ and CH$_4$) indicating a limited (<6 vol.%) amount of corresponding clathrates in plume source regions. Another problem of this model is that water vapour is created by sublimation of a major part of these ice particles which are initially entrained in the flow of gases released from clathrates. The formation of sodium free water vapour (Schneider et al. 2009) through sublimation of icy grains in clathrate-decomposition models is not consistent with the dominance of salt rich ice grains. Either some sodium would enter the gas phase through grain sublimation or – if that can be avoided – almost pure salt grains would be left behind in the gas flow which is also not observed.

In contrast to the composition of ice grains the composition of plume gases does not provide a strong constraint on liquid water as the major plume source (despite the stability of NH$_3$/H$_2$O-rich solutions in the south polar ices). If plume gases exclusively originate from a liquid water reservoir, low-solubility species would be more abundant than high-solubility compounds. However, the most reliable data show no clear correlation with solubilities in water (Fig. 10). On the one hand, highly soluble methanol looks strongly depleted compared to cometary abundances. On the other hand, highly soluble NH$_3$ and shows no signs of depletion, whereas low-solubility CO, N$_2$, CH$_4$, C$_2$H$_2$, C$_2$H$_6$, and other hydrocarbons are less abundant or not detected in the plume. The lack of a clear correlation of INMS data with solubilities (Fig. 11) is not consistent with degassing solely from a subsurface liquid water reservoir. A low (far from eutectic) NH$_3$/H$_2$O ratio in the plume implies a contribution from dry sources, though degassing of NH$_3$-rich brine pockets together with surrounding ices is possible. The detection of NaHCO$_3$ and/or Na$_2$CO$_3$ in icy grains (Postberg et al. 2009) indicates the presence of HCO$_3^-$ and CO$_3^{2-}$ ions in the solution, and some CO$_2$ degassing could be driven by changes in solution pressure, temperature, or pH (Matson et al. 2012).

The lack of abundant rocky solids in plume emissions put additional constrains on the phase composition of plume sources. The dominance of ice in plume solid emissions is consistent with separation of rocks during differentiation and does not indicate a presence of cometary-type rock-ice mixtures in the plume source region.
The deficiency of rocks remains to be reconciled with the presence of CH$_4$, NH$_3$, among plume gases. One possible explanation is the heterogeneous accretion of Enceladus (Charnoz et al. 2011, section 3.3.1) that led to preferential accumulation of solid volatile species (ices, clathrates, hydrates) in the peripheral parts of the moon. Current tidal warming of the south polar region could cause both melting of volatile-poor salty ices and release of gases from never melted volatile-rich ices.

Although the composition, grain size and dynamics of the emitted ice particles strongly point at a source of salt-bearing aqueous solution it is not unequivocally proven. The composition and abundance of volatiles in the gas plume is not in agreement to be solely emitted by evaporation of water. However, all currently proposed dry scenarios encounter conflicts with one or more of the plume observations. A mixture of liquid and solid sources contributing to the plume is the likely scenario. Whereas the liquid sources would be responsible for most of the ice emission, ice sublimation and possible clathrate decomposition have a significant contribution to the gas phase.

### 3.3 Chemical evolution scenarios

#### 3.3.1 Accreted materials and chemical effects of water-rock differentiation

Enceladus could have accreted at low temperatures and pressures from either a slow-inflow accretion disc in the solar nebula (Canup and Ward 2002, 2009) or from massive rings of Saturn formed significantly later after the formation of the planet (e.g., Porco et al. 2007, Charnoz et al. 2010, 2011, Canup 2010). Enceladus was probably accreted as a mixture of ices, condensed organic compounds and volatiles, and rocky components. Water ice dominated over other ices, and other volatiles could have accreted as hydrates (NH$_3$, HCl), ices (CO$_2$, NH$_3$, CH$_3$OH), and clathrates (CO$_2$, H$_2$S, CH$_4$, and some other light hydrocarbons) (e.g., Hersant et al. 2008). Although the nature of accreted organic matter is unclear, some material could have resembled high molecular weight organics common in carbonaceous chondrites and C-H-O-N particles detected in the comet Halley dust (Jessberger et al. 1988, Fomenkova et al. 1994). The rocky component was presented by anhydrous and reduced minerals and amorphous silicates, by analogy with primitive materials in chondrites (e.g., Brearley and Jones 1998) and cometary dust (Jessberger 1999, Hanner and Bradley 2004, Zolensky et al. 2006). This component may not have abundant chondrules. The minerals could be dominated by enstatite (MgSiO$_3$), forsterite (Mg$_2$SiO$_4$), Fe-Ni-metal, and troilite (FeS).

If short-lived $^{26}$Al was abundant in the rocks at the time of accretion, melting of ices could have caused water-rock differentiation within $\sim 10^6$ years (Schubert...
Enceladus as an active body

Hypothetic scenarios that included enhanced heat sources on early Enceladus (due to abundant $^{26}$Al and large tidal dissipation) may have involved hydrothermal processes, some dehydration of earlier hydrated phases, intense alteration and oxidation of organic species, oxidation of NH$_3$ and N$_2$ production (Matson et al. 2007; Glein et al. 2008). These pathways also suggest a significant thinning of the primordial ice-bearing shell (Schubert et al. 2007). However, $^{26}$Al could have significantly contributed to the thermal budget only if the satellite accreted within a few millions years after the formation of the CAIs (the first major solids formed in the solar nebula). This early accretion is difficult to reconcile with the present knowledge of the Saturn system formation (e.g., Sasai et al. 2010) and with the possible formation of mid-size satellite from Saturn rings (e.g., Charnoz et al. 2011). Typical cometary abundances of NH$_3$ and CH$_4$ in plume gases (Fig. 11), the low limit for N$_2$ in plume gases (Hansen et al. 2011), and the low K/Na ratio in E-ring particles (Fig. 12) also indicate moderate heating without major hydrothermal activity in Enceladus’ history. Without abundant $^{26}$Al, the differentiation could have been driven by the decay of long-lived radionuclides and occurred about 0.5–1 Gyr after accretion (Schubert et al. 2007; Malamud and Prialnik 2013). The release of heat in mineral hydration reactions (e.g., Grimm and McSween, 1989; Malamud and Prialnik, 2013) contributed to water-rock differentiation. Tidal dissipation in the warmed interior may have also accelerated the differentiation process (Schubert et al. 2007).

Independent of the source of heat, temperature rose more rapidly in the central part of the body and led to the melting of ice, resulting in the gravitational separation of rocky grains and melts, which formed a growing liquid layer above a rising rocky core (Schubert et al. 2007). Initial low-temperature fluids were rich in species that strongly decreased the melting temperature of ice (NH$_3$, methanol, Cl$^-$) and became enriched in metals along with warming of fluids and dissolution of minerals. Release of volatiles from melting ices led to diverse chemical transformations that included dissolution, hydrolysis, and other chemical interactions. As an example, dissolution of C oxides could have led to carbonate, bicarbonate, and formate (HCOO$^-$) anions. Hydrolysis of ammonia produces the ammonium ion (NH$_4^+$) which reacts with anions leading to precipitation of NH$_4$Cl and ammonium carbonates/bicarbonates (Kargel 1992; Marion et al. 2012). These reactions should have diminished the concentration of NH$_4^+$ in solution. Hydrolysis of formaldehyde causes its polymerization to methylene glycol, CH$_3$(OH)$_2$. Hydrogenation of carbon species in H$_2$-rich environments (see next paragraph) may lead to formation of formate, methanol, CH$_4$, and other hydrocarbons. These and other chemical transformations should have changed the abundances of accreted volatile compounds and led to new species that accumulated in solutions and/or partitioned through precipitation or liquid phase separation. Some of these species could have been sampled by the Cassini INMS and CDA instruments, especially in association with salt-bearing grains.

Aqueous alteration of rocks starts with partial and unequal dissolution of solids followed by precipitation of secondary phases. Water is mainly consumed in hydration and oxidation reactions. As in chondritic materials, oxidation could involve formation of magnetite, Fe-bearing olivine and phyllosilicates from Fe-Ni metal, tochilinite from troilite, phosphates from phosphides, and Ni-sulfides from Fe-Ni metal. Hydration leads to formation of phyllosilicates (serpentine, cronstedtite, saponite, chloride) and tochilinite. Precipitation of carbonates partially consumes inorganic carbon and cations (Ca$^{2+}$, Mg$^{2+}$, Fe$^{2+}$, NH$_4^+$) from solution. Fluid composition and pH becomes strongly affected by the solubility of newly-formed minerals. Evolution of fluid chemistry leads to an alkaline NaCl solution with subordinate amounts of carbonate and bicarbonate ions, and K$^+$ (Zolotov 2007, 2012). Although Fe, Ni, Mg, Ca, Si, Al, and S are abundant in the rocks, they are minor constituents of solutions owing to low solubility of corresponding secondary minerals (mainly phyllosilicates and sulfides). Geologically rapid water-rock reactions during differentiation could have produced a NaCl- and NH$_3$-bearing solution layer above the core. The solution composition should not be much different if the differentiation occulted without formation of a large global water layer.
Subsequent exhaustion of heat sources caused freezing of fluids and precipitation of salts, especially around the rocky core. Methane and CO$_2$ could be trapped in clathrates in most of the new-formed ice (section 2.2.2). Although CO$_2$ clathrate is more stable, conversion of CO$_2$ to HCO$_3^-$ and CO$_3^{2-}$ ions in alkaline fluids lowered CO$_2$ fugacity and constrained clathrate formation.

The composition of plume gases indicates survival of some accreted icy material during differentiation. This is consistent with the thermal evolution models of Schubert et al. (2007) that do not imply melting of a primordial icy-rocky shell of accreted materials. However, the lack of detection of rocky particles by the Cassini CDA instrument is not consistent with gas origins in remnants of undifferentiated crust. In addition, the crater record (section 2.1.5) and the composition of the Enceladus’ surface and plume solids do not indicate occurrence of rock-bearing materials in the upper surface layers. These observations could be explained by heterogeneous accretion (Charnoz et al. 2011), a gravitational submergence of the initial shell, and cryovolcanic activity within the primordial shell. In a case of heterogeneous accretion, a low-density volatile-rich icy shell could have floated above an early ocean. In a case of early homogeneous accretion, a strong heating by $^{26}$Al caused thinning (Schubert et al. 2007) and submergence the rock-bearing shell followed by its partial or complete melting.

Cryovolcanic activity could have been driven by disruptions in the primordial shell caused by impacts, tidal motions, and volume changes due to ocean freezing (e.g., Manga and Wang (2007)) and hydration/dehydration processes in the core. The non-volatile chemistry of rapidly frozen dikes, cryovolcanic flows at the surface, and ancient plume deposits would represent the composition of aqueous fluids. However, low-solubility gases (CO, N$_2$, H$_2$, CH$_4$ and other hydrocarbons) could have been lost from dikes, cryovolcanic flows, and eruptive vents. Decompression also leads to some CO$_2$ degassing through shifting of the carbonate equilibria (CO$_2^−$ → HCO$_3^+$ → CO$_3^{2-}$) in alkaline NaCl-NaCO$_3$/NaHCO$_3$ solutions suggested in the interior (Postberg et al. 2009; Zolotov 2007). In deep dikes and other intrusions of aqueous fluids, CH$_4$ and CO$_2$ could be trapped in clathrates together with salts precipitated upon freezing within the primordial icy-rocky shell. Other oceanic impurities (NH$_3$, methanol, and other organic species), organic liquid phases, and low-solubility gases (H$_2$, CO) could be trapped in gas/liquid inclusions in ice. Ices formed through cryovolcanism should be compositionally distinct from accreted ices. It follows that the comet-like concentrations of CH$_4$ and NH$_3$ in plume gases are not consistent with gas origin in secondary ices.

### 3.3.2 Chemical processes related to tidal heating

The presence of salts, clathrate hydrates, ammonia and/or methanol in water ice is known to affect the rheological properties of the mixture (Arakawa and Maeno 1994; Cole et al. 1998; De La Chapelle et al. 1999; Durham et al. 2003, 2005, 2010; McCarthy et al. 2011). As a consequence, their presence in the icy shell is expected to modify the response to the tidal forcing. At temperature near the melting point, the presence of salts, ammonia, and methanol favor the formation of melt pockets at grain boundaries, owing to their anti-freezing properties (e.g., Fortes and Choukroun (2010)). Experimental tests on water ice containing brine pockets (De La Chapelle et al. 1999) indicates that a melt fraction of about 5% decreases the viscosity of ice at the melting point by about one order of magnitude (Tobie et al. 2003). Cyclic loading experiments on terrestrial sea ice (Cole et al. 1998) indicate that the viscous strain rate increases markedly with brine volume, suggesting that tidal dissipation should be significantly enhanced in regions enriched in brine pockets. Brine pockets could, for instance, be entrained by convective motions in the icy shell and be concentrated in warm upwellings (Zolotov et al. 2004; Pappalardo and Barr 2004).

Ammonia is expected to strongly impact the tidal dissipation process owing to its very strong effect on the melting temperature. Laboratory experiments show, indeed, that partially melted ice containing 4% of ammonia would have a viscosity ranging between $10^9$ and $10^{11}$ Pa s at temperatures 180-210 K (Arakawa and
Maeno 1994), whereas the viscosity of pure water ice at the same temperature
would be at least 10^{20} \text{ Pa s}. A slush layer enriched in ammonia may form sub-
sequently to the downward freezing of an internal ocean, since the concentration
of ammonia progressively increases in the remaining liquid. An isoviscous slush
layer of 10^{10} \text{ Pa s} could, for instance, produce a total dissipation power of about
2 \text{ GW} (Tobie et al 2008). Likewise, tidal dissipation is expected in a case of salty
brines and/or organic liquids accumulated at the core-icy shell interface (Zolotov
2007). Dikes of enriched inorganic and organic liquids that may be injected in
the icy shell during crystallization episodes (see section III.2.2) should also locally
enhance the tidal dissipation rate. This will help strike-slip motions between differ-
ent icy blocks (e.g. Nimmo et al (2007)). Local tidal dissipation would cause
further melting which, in turn, induces dissipation. However, significant melting
would limit tidal dissipation. Further modeling is needed to understand the com-
plex coupling between anti-freezing compounds, tidal dissipation, and convective
motions.

At temperatures far below the melting point of pure water ice, the presence of
ice impurities (clathrates, salts or silicate microparticles) usually tends to make
the ice mixture more resistant to creep (e.g. Durham et al (2003, 2005); McCarthy
et al (2011)). However, mixtures of water ice and hydrated salts appear to be more
brittle than pure ice (McCarthy et al 2011). This may favor the formation of a
zone of weakness promote tidal deformation in the icy shell. Even small amounts
of impurities also have an effect on the evolution of the ice grain size (e.g. Durand
et al 2009). In polar ice sheets on Earth, for instance, soluble and second-phase
particles located at grain boundaries affect grain growth by exerting a pinning force
on moving grain boundaries (e.g. Alley et al 1986). As the deformation of ice is
sensitive to grain size in the stress conditions expected within Enceladus’ ice shell
(Barr and McKinnon 2007), the impurities, even if present in small concentrations,
could have a major impact on the rheology of ice. Both experimental investigations
and numerical modeling are required to understand the interplay between solid
impurities, rheology, and tidal dissipation.

3.3.3 Habitability of Enceladus

The likely occurrence of aqueous fluids, organic species, and the detection of carbon
species with diverse oxidation states (CO\textsubscript{2}, CH\textsubscript{3}OH, CH\textsubscript{4}) imply habitable envi-
ronments in the interior of Enceladus. Although the composition of salt-bearing
particles emitted from Enceladus (Postberg et al 2009, 2011) indicates present
or past aqueous environments it does not provide direct indication for life. The
composition of plume gases does not indicate their aqueous sources, though the
deeper interior could contain altered species transformed in abiotic (section 3.3.1)
or biologic processes. Likewise, the comet-like isotopic ratio of H in plume gases
(Waite et al 2009) may not characterize deep water-bearing environments.

If life exists, it may not be limited by the ability of C-N-O-H-bearing nutri-
ents and biologically-important metals, though the low solubility of phosphates
in alkaline fluids may confine biological productivity. The moderately alkaline pH
evaluated for Enceladus’ fluids (Postberg et al 2009; Zolotov 2007) is suitable for
known microorganisms. However, organisms may not live in extremely cold and
alkaline NH\textsubscript{3}-rich fluids (at 173 to ~230 K) that are characterized by low activity
of water.

Putative life on Enceladus could be presented by chemotrophic microorganisms
that gain metabolic energy from oxidation-reduction reactions (McKay et al 2008
Parkinson et al 2008). Major metabolic pathways could involve reduction of C
species (CO\textsubscript{2}, CO, formaldehyde, methanol, and unsaturated hydrocarbons) by
H\textsubscript{2} produced through water-rock interactions (McKay et al 2008). Methane is
the final product of these reactions. Aqueous hydrogenations of formaldehyde,
and unsaturated hydrocarbons to produce methanol are energetically favorable
as well. However, traces of O\textsubscript{2} produced through radiolysis of ice or liquid water
(e.g., via radioactive decay of \textsuperscript{40}K, Chyba and Hand (2001)) may not be used by
organisms that oxidize H\textsubscript{2}, C, S, and Fe\textsuperscript{2+} species because of rapid consumption of
O₂ through oxidation of sulfides to native S and/or sulfates. The abiotic formation of some sulfates creates a potential for biological sulfate reduction by H₂, CH₄, and other organic species [McKay et al. 2008]. However, a large-scale abiotic oxidation of sulfides may not have occurred in cold and/or short-living aqueous environments with a limited recycling of surface oxidants. Therefore, the mass of sulfate available for biological reduction could be negligible.

In the history of the moon, redox disequilibria could have been activated through a melting of volatile-rich ices following by water-rock-organic interactions. Organisms may have adapted to changes of prevailing redox disequilibria and evolving water activity. The organisms could have been able to survive (in a dormant stage) in sub-freezing environments without aqueous fluids and in near-eutectic brines rich in ammonia, chlorides, and organic liquids at temperatures >160-170 K. Signs of life could be potentially detected in plume components that indicate aqueous origins (e.g., salty ice grains).

3.4 Summary

The surface materials and plume emissions are mainly presented by water (crystalline and amorphous ice and vapor). Enceladus’ emissions also contain CO₂, NH₃, CH₄, CH₃OH, organic species with mass >100 Dalton, chlorides and carbonates/bicarbonates of Na and K, and nanometer-sized silica grains.

Although the composition, size and dynamics of the emitted ice particles strongly point at a source of liquid salty water, there is no unequivocal chemical evidence for a large water reservoir that supply the plume components. Solution pockets enriched in ammonia are stable in the warm water ice at the surface but the composition of plume gases is not consistent with degassing of aqueous fluids and implies a contribution from sublimation of heated ice that contains volatiles in clathrates, brine pockets, or structural defects. The composition of emitted gas end solids points at a diversity of processes creating the plume. A mixture of liquid and solid sources is the most likely scenario. A south polar low-viscosity layer between the rocky core and the ice shell is likely although it is unclear how it would connect to possible saltwater pockets at shallow depth feeding the plumes. The possible occurrence of primordial newer melted volatiles together with aqueously-processed compounds in the icy shell suggests incomplete differentiation of the moon and may indicate spatial separation of these diverse species. The compositional heterogeneity should reflect mechanisms of water-rock differentiation and subsequent thermal processes in the icy shell. Chemical data indirectly indicate predominantly low-temperature (≤273 K) water-rock interaction in Enceladus’ history. This implies a low amount of accreted ²⁶Al and a late, possibly heterogeneous accretion is consistent with the current formation models for the Saturn system.

Although Enceladus could be habitable and may harbor chemotropic organisms (e.g., methanogens) in subsurface aqueous environments, current compositional data do not indicate biological processes.

4 Plume formation and properties

The Enceladus activity is signified by the plume towering the south pole of the satellite. The plume allowed for a direct in situ sampling of gas and particles during Cassini flybys, providing unique information on the composition of material stemming from the interior of the moon. Moreover, the physical characteristics of the plume pose important constraints on the activity in general, such as the production rates of vapor and particles and their variability, the ejection speeds, the particle size distribution, and the distribution of plume sources and their relation to features and temperatures on the surface.
Fig. 13 Left panel: Cassini image PIA08386 (NASA/JPL, http://www.ciclops.org), showing a high phase image of the Enceladus plume. Visible are numerous dust jets but also a broad distribution of dust. Right panel: Cassini image PIA11688 (NASA/JPL, http://www.ciclops.org). A close-up view of the south polar terrain at large phase angle, bringing out details of the dust ejection. In this particular viewing geometry the terminator crosses the Tiger Stripes. Dust emerging from those sources on the dark side of the shadow becomes visible at height when leaving the shadow. Arrows mark the shadow boundary, which becomes visible because of the presence of dust.

4.1 The gas and dust components of the plume

The dust plume is directly seen in the high phase images from the Cassini ISS [Porco et al 2006; Spitale and Porco 2007; Ingersoll and Ewald 2011], in the near infrared data from Cassini VIMS (Hedman et al 2009; Hedman et al 2013) and it has been recorded in situ by the Cassini CDA (Spahn et al 2006b; Schmidt et al 2008; Kempf et al 2010; Postberg et al 2011). This part of the dust configuration is formed by grains composed dominantly of water ice (Postberg et al 2011) and sized in the sub-micron to micron range (Spahn et al 2006b; Hedman et al 2009; Postberg et al 2011). Additionally, jets of very fine, charged particles were detected by the Cassini CAPS instrument (Jones et al 2009; Hill et al 2012). The properties of these jets are discussed in detail in section 5.2.3 and we focus in this section primarily on the larger grains.

The sources of the dust plume are the Tiger Stripes, with a spatially variable level of activity. Several prominent sources were identified with particularly hot regions on the Tiger Stripes (Spitale and Porco 2007). A later, preliminary analysis of images reports a much larger set of sources (Porco et al 2011). The distribution of gas sources is perhaps best constrained from stellar and solar occultations by the plume recorded with the Cassini UVIS experiment. In a similar manner as for the dust plume, the UVIS data shows evidence for a broad ejection of gas with additional peaks in gas density, which were shown to be correlated (Hansen et al 2008, 2011) with the most prominent dust jets (Spitale and Porco 2007).

The gas and the dust phases of the south-polar ejection exhibit important differences, though, in spite of the fact that they likely originate from the same sources. Generally, the gas ejection speeds are found to be on the order of several times the escape speed (≈ 237 m/s) [Hansen et al 2006; Tian et al 2007; Smith et al 2010; Dong et al 2011] while the scale height of the dust plume suggests mean ejection velocities on the order of 100 m/s only [Porco et al 2006; Schmidt et al 2008; Hedman et al 2009]. This dynamical difference has important implications. Namely, it shows that there cannot exist a strong momentum transfer from the dust to the gas, simply because the gas would then be efficiently slowed down by the particles, which is not seen. On the other hand, for plausible subsurface gas densities [Schmidt et al 2008] one expects a strong momentum transfer from the gas to the dust, mediated by friction [Epstein 1924; Cunningham 1910]. One possibility is that particles are decelerated relative to the gas flow already in the subsurface vents. This has been explored in detail by Schmidt et al (2008), who suggest that particles, entrained in the subsurface gas stream are repeatedly decelerated by collisions with the walls of the vents. Another principal possibility is that both gas
and dust emerge at slow, subsonic speed from the vents and the gas is accelerated further by pressure forces, when expanding above the vents, and before the gas becomes collisionless. The latter scenario has not been explored to date and it is not clear if this is realistic for the observed gas and dust densities, as well as the distribution of velocities and grain sizes.

### 4.2 Constraints

The observed properties of the gas and dust plumes pose important constraints on possible scenarios of plume formation. In this section we summarize such constraints and discuss models suggested in the literature in this light.

#### 4.2.1 Steadiness and variability

In data from stellar and solar occultations by the plume observed by UVIS there is little evidence for a variability of the vapor production rate (Hansen et al. 2011) within the error limits of the data reduction. Especially, no variation was seen in correlation with the orbital phase of Enceladus, which was predicted as a consequence of the tidal flexing of the south polar terrain and the Tiger Stripes (Hurford et al. 2007), although the coverage of orbital longitudes with occultations was limited. Saur et al. (2008) modeled the influence of the water plume on the corotational plasma and derived the corresponding magnetic field perturbation. From comparison to data obtained by the Cassini MAG instrument at multiple flybys they infer plume source rates that exhibit a significant level of variability (see Tab. 3). Dong et al. (2011) fitted a model of eight water vapor sources, identified with the jet sources by Spitale and Porco (2007), to measurements by the INMS instrument recorded during the E3, E5, and E7 encounters. They found ejection speeds and plume densities consistent with those inferred by UVIS. In contrast, Smith et al. (2010) infer a variability by a factor of four between E3 and E5 (seven months).

In a recent analysis of VIMS data Hedman et al. (2013) find a periodic variation of the dust plumes’ brightness that correlates well with the orbital phase of Enceladus (i.e. with a period of \(\approx 30\) hours). This supports well earlier ideas on the influence of tidally induced normal stress on the width of the Tiger Stripe cracks (Hurford et al. 2007), possibly with a phase shift between forcing and response due to the effect of a physical libration of Enceladus (Hurford et al. 2009). Such a tidally controlled periodicity of the Enceladus activity has consequences for the interpretation of data and modeling, which need to be explored in future investigation. The observed fast modulation of the plume brightness is primarily a consequence of the variation of the cross section of the vents over the tidal cycle, leading to a corresponding variation of the gas and dust fluxes. It is not expected that the underlying subsurface distribution of heat varies noticeably on the same timescale.

Besides this periodic brightness variation with the tidal cycle the plume appears to be continuously active at least over the time span of the Cassini observations and the question arises how long has it been active in the past. One lower limit on the duration of the activity comes from the lifetimes of the E ring particles, because the plume must sustain this ring. These lifetimes vary strongly with particle size (Kempf et al. 2010), but it seems that the directed, vertical injection to the E ring of particles through the south polar plume tends to increase the lifetimes of particles (Juhász et al. 2007) when compared to pre-Cassini models of the ring (Juhász and Horányi 2002; Juhász 2004). Overall, it should take roughly 100 years to build up a steady E ring. But most probably the timescales for plume activity rather correspond to the much longer geological timescales, comparable to the ages inferred for the south polar terrain itself, which is on the order of \(10^7\) years or less (Porco et al. 2006).

One age constraint was derived from certain features on the Enceladus surface seen in images, which were interpreted as deposits of plume particles (Schenk et al. 2011). The comparison of the estimated deposit thickness to accumulation rates
Enceladus as an active body

obtained for the same location from ballistic particle models (Kempf et al 2010) yielded a duration of 10s of millions of years for the plume activity. However, this preliminary analysis needs to be extended to a larger set of images, from multiple locations on the surface, so that the effect can be unambiguously attributed to the action of plume particles falling back onto the surface.

4.2.2 Dust to gas mass ratio

One important constraint for models of plume formation is the dust to gas ratio. The total dust mass in the plume was derived by Ingersoll and Ewald (2011) from photometry of the plume and an ice to vapor ratio of 35 to 70%. At least the upper limit appears uncomfortably large for pure condensation models, even if one takes into account the afore mentioned effects. For instance, from models invoking condensation as the process of particle creation one finds that typically about 5 to maximally 10% of the gas discharge can be condensed (Schmidt et al 2008), depending on the initial enthalpy of the gas and the level of supersaturation reached in the flow. On the other hand, from these numbers follow significantly larger dust to gas ratios in the plume along a given line of sight. In practice, such line of sight densities are inferred for the gas from occultations and from the local brightness of the dust plume. When comparing to the bare source rates one needs to take into account that the dust is ejected at slower speeds, thus contributing to the density in a given volume element of the plume more efficiently than the gas. Also, the gas is faster than the escape speed, and thus, each molecule traverses the plume only once before it escapes the satellite. In contrast, most of the dust grains fall back, so that they traverse the plume twice before they re-impact the surface.

An additional contribution to the plume mass may come from particles that are entrained in the gas flow as frozen droplets from a spray above a liquid reservoir, as it was proposed by Postberg et al (2009, 2011). In this case only a part of the particulate mass would condense. The details of this model (energy budget, size distribution of grains) have not been worked out to date, but the concept might be consistent with the model proposed by Matson et al (2012), to explain the subsurface transport of heat and material on Enceladus by a cycle driven by the exsolution of gases. Bubbles produced by such gases could generate spray above the liquid, with droplets that have a composition identical to the liquid.

4.2.3 Grain size and velocity distribution

Particle sizes in the plume were derived in situ by Cassini CDA (Spahn et al 2006). For the flyby E2, which traversed relatively distant parts of the plume, they inferred a size slope of $\approx -3$ for the differential size distribution. In the E ring radially close to Enceladus (Kempf et al 2008) inferred steeper differential size slopes between $-4$ and $-5$, indicating that smaller particles have a higher probability to escape the moon’s gravity field. From dust impacts on the Cassini RPWS instrument a yet steeper distribution with a power law exponent around $-6$ was inferred (Kurth et al 2006). The likely explanation for the different exponents is that the true distribuiton of grains cannot be approximated well as a power law in the full range of particle sizes covered by various instruments. Probably the distribution falls of more rapidly than a power law and CDA fits exponents to a flatter part of the distribution for sizes from sub-micron to about two microns, whereas RPWS sees the steeper distribution in the few micron range.

For the velocity distribution of particles leaving the Enceladus vents Schmidt et al (2008) derived the form

$$ f_R(u) \, du = \frac{R}{R_c} \left( 1 + \frac{R}{R_c} \right) \frac{u}{u_{gas}} \left( 1 - \frac{u}{u_{gas}} \right)^{1 - \frac{R}{R_c}} du, \quad (3) $$

where $f_R(u)$ is the differential velocity distribution that depends parametrically on the particle radius $R$. The critical radius $R_c$ lies in the submicron range and it
depends on the dynamical coupling between gas and grains. Here, \( u_{gas} \) is the gas velocity. A comparison to VIMS data by Hedman et al (2009) suggested \( R_c \approx 0.4 \mu \text{m} \).

Formula (3) follows from the statistics of particles that are decelerated repeatedly by collisions with the walls of the vents and then are re-accelerated by gas friction. It can, however, more generally be regarded as a convenient parameterization for the size-dependent speed distribution. The distribution is normalized to unity in

\[
0 \leq u \leq u_{gas}
\]

Then formula (3) implies that grains with sizes \( R > R_c \) have a speed distribution that peaks at a velocity between 0 and \( u_{gas} \), while grains with sizes \( R < R_c \) have a distribution with one single pole at \( u = u_{gas} \).

Using equation (3) one can construct the combined distribution for particle sizes and ejection speeds

\[
f(R, u) = f_R(u) p(R)
\]

by multiplication with a suitable size distribution \( p(R) \). For instance, Postberg et al (2011) used a power law for \( p(R) \) to construct from equation (4) a plume model with a size distribution that varied with position in the plume (see also section 3.1.3 and figure 7).

From the analysis of VIMS images taken from the Enceladus plume at high phase angles, with a high spectral resolution in the near infrared, Hedman et al (2009) reconstructed the size distribution in the plume and its variation with altitude above the south pole. They find that generally the sub-populations of larger particles have smaller scale hights than the small ones. The size distributions at lower altitudes are found to be shallower than a power law with exponent – in the range from 1 to 2 microns but are steeper than such a power law at larger size. Generally, a fairly steep drop is observed in the particle abundance for sizes larger than 3 micron. Hedman et al (2009) assumed that the dust density in their measurement is dominated by those grains reaching their maximal altitude where their line of sight intersects the plume. Using the integrals from the two body probelm they could then invert their data to reconstruct a distribution of starting velocities. The size-dependence of the slopes of the so determined velocity distribution is roughly consistent with formula (3).

### 4.2.4 Supply rate of particles to the E ring

Juhász and Horányi (2002) derived an E ring model with particles generated at various satellites in the E ring. The evolution of the particles was subject to Saturn’s gravity (including higher order contributions due to the oblate shape), solar radiation pressure, and electrodynamic forces, and plasma drag, with an appropriate charging model to obtain the instantaneous grain charge in the magnetosphere. The sinks of the particles are collisions with Saturn’s A ring, the E ring moons and plasma sputtering of the grains. This model did not take into account the directed injection of particles from Enceladus through the south polar plume. A steady state E ring was obtained at a production rate of about 1 kg/s particles at Enceladus.

Spahn et al (2006b) fitted a model of particles ejected from Enceladus’ south polar terrain to CDA data from the flyby E2. Their best match with the model suggested a supply rate of 0.2 to 1 kg/s from Enceladus to the ring. Porco et al (2006) derived from the analysis of Cassini images an escape rate of 0.04 kg/s.

Schmidt et al (2008) constructed a model plume based on equation (4) and particle size-distributions obtained condensation in the subsurface vents. Parameters of the model were constrained from fits to CDA data from E2 and to brightness profiles of images. They obtained a mass production rate of 5kg/s from which 0.5 kg/s escape to the E ring. From high phase images of the plume Ingersoll and Ewald (2011) derived an about 10-fold value for both the production and the escape rate.

### 4.2.5 Composition: Sodium and volatiles

The composition of the water plume has been derived from INMS (Waite et al 2006, 2009, 2011) and UVIS data (Hansen et al 2006, 2008, 2011). The current
interpretation of these data is summarized in section 3.1.2 and the implication for
liquid and solid sources of the gas are discussed in section 3.2.4. The ice grains in
the E ring and in the plume (Postberg et al. 2009, 2011) were found to split into
different compositional types (see section 3.1.3). Importantly, most of the grains
show at least small amounts of sodium which can be attributed unambiguously to
the presence of salt. Moreover, (Postberg et al. 2011) have shown that most of the
ice mass in the dust plume has a salinity on the order of 1%. On the other hand,
from earth-bound observation (Schneider et al. 2009) showed that the gas in the
plume is practically free of sodium, from analysis of the brightness of the plume
at the wavelength of the sodium D2 line.

4.3 Scenarios for plume formation

4.3.1 Explosive boiling model

This scenario for plume formation was proposed by (Porco et al. 2006). It assumes
that liquid is present very close to the surface (motivated by the high surface tem-
peratures seen by CIRS (Spencer et al. 2006)) and that fractures, rapidly opening
in the ice, suddenly expose the liquid water to vacuum. In this zero pressure
condition the uppermost layer of liquid should instantaneously begin to boil while
simultaneously part of it begins to freeze. Part of the frozen phase would be ejected
as plume particles, accelerated and dragged to space by the expanding gas. This
process occurs far from equilibrium, and thus, in principle large solid to gas ratios
should be possible. But the process is also self-limiting, because any crack should
be rapidly sealed by the freezing part of the liquid. This is hard to reconcile with
the ongoing quasi-continuous activity, as is observed, even if one invokes multiple
cracks to maintain an average level of activity. Moreover, in this fast process the
gas from the boiling phase, as well as the frozen part should have the same sodium
content as the liquid from which they originate. But this is not observed: the plume
vapor is essentially sodium free (Schneider et al. 2009) compared to the 1% salin-
ity of the particles (Postberg et al. 2011). Brilliantov et al. (2008) have estimated
for the velocity of the expanding gas from basic thermodynamic principles. They
found that the maximum speeds should be as low as meters per second, which is
not consistent with observations, implying gas velocities on the order of several
hundred meters per second or higher (Hansen et al. 2008).

4.3.2 The clathrate reservoir hypothesis

Kieffer et al. (2006) proposed a dry model for the plume activity that is based on
the decomposition of gas hydrates (clathrates) in a subsurface reservoir (see also
Gioia et al. (2007a)). This is motivated by the observation that plume gases like
CH4 and CO2 can form clathrates, and then exhibit higher solubility in water
ice than in liquid. These clathrates become unstable at high temperatures and/or
low pressures so that near cracks, or heat sources, they would decompose into
(volatile) gas and solid ice. If there exists a shell of clathrate-rich ice below the
south-polar terrain, a decomposition front could propagate from the Tiger Stripes
into this reservoir. The advection of the gases to the surface through the Tiger
Stripe faults would explain the volatile gases found in the plume. On the other
hand, the process does not produce water gas directly, which would need to form
under subsurface conditions by sublimation from the icy break-up products of the
clathrates. This might be difficult to achieve thermodynamically, especially at the
relatively low temperatures for which the model, also called “frigid faithful” was
proposed. The pressure of the gas should rapidly adjust to the saturation vapor
pressure, which is a strongly dropping function of temperature. The most severe
difficulty of this model, as for any dry scenario, is that it cannot account for the
observed sodium in the plume grains (Postberg et al. 2011).
4.3.3 Shear heating models and sublimation from warm ice

The production of heat by tidally induced shearing motion in the south polar ice cap was modelled by Nimmo et al. (2007). The frictional heating results from strike-slip motion along faults (the Tiger Stripes) in the upper, brittle ice layer and from viscous heating in the ductile ice at depth. The mechanism is efficient if there exists an ocean between the rocky core of the satellite and the ice mantle, so that the tidal displacements of the latter can be large. The heat produced in the ice can be transported by conduction, by the hydrodynamic flow of vapor through the cracks, and by diffusion of vapor if the ice is porous. Nimmo et al. (2007) considered a balance of non-equilibrium vapor production and diffusion while Ingersoll and Pankine (2010) suggested to limit the vapor density so that it locally adjusts to the saturation vapor pressure (see also discussion in section 2.2.4).

As a result, one expects the ice to warm up along a relatively thin zone that extends from the Tiger Stripes downward into the ice cap. The process reaches an equilibrium temperature distribution as depicted in figure (14), resulting from energy input in the shear zone, redistribution of heat by conduction, vapor diffusion and advection, and loss due to radiation at the surface and vapor escaping to vacuum. Additionally, heat may be effectively transported in form of latent heat of the vapor. It can be transferred to the ice when vapor re-condenses on the walls in the upper part of the faults. In total, on the surface this would result in a heat flow pattern that is localized at the Tiger Stripes (see Fig. 3c).

This process is capable of producing abundantly water vapor, possibly reaching the measured mass fluxes, if the temperatures reach sufficiently high values, depending on the shear rate, the unknown thickness of the ice and poorly known parameters, as the viscosity of the ice. But the model cannot account directly for the observed non-water gases. To this end, an additional process must be invoked, like outgasing of volatiles contained in the ice, similar to the clathrate decomposition model. The particles of the dust plume might be created in this scenario by condensation in the vents as proposed by Schmidt et al. (2008). However, the model cannot account for the presence of sodium salts in the ice particles.

4.3.4 Evaporation from liquid

Conditions for melting in the shear heating model were investigated by Ingersoll and Pankine (2010). One main conclusion from that work was that melting can indeed occur even if the heating process is self-limiting because of the temperature dependence of the viscosity of the ice. However, the precise conditions would
Enceladus as an active body

**Fig. 15** Figure reproduced from Postberg et al. (2009). Formation of salt rich droplets by dispersion from the liquid (a) which subsequently freeze and grow to some extent by condensation of vapor. Salt poor grains could arise directly from homogeneous nucleation in the vents (b).

depend on the unknown grain size in the ice (because this determines the porosity and thus the magnitude of vapor diffusion) and the width of the cracks (because this would limit the amount of heat advected with the hydrodynamic vapor flow).

If the melting affects brine pockets in the subsurface ice, then there is no principal problem with having sodium in the grains. Such frozen brine pockets may result from previous phases of activity in the Enceladus history in a scenario of an intermittent activity cycle (Roberts and Nimmo 2008), and they might be part of a convection system in the ice shell (Tobie et al. 2008) (see also Fig. 2).

The liquid scenario offers a simple explanation for the detection of sodium salts in the ice grains, if the liquid reservoir is, or was in contact with the rocky core. Enceladus, enriching the water with minerals. In fact, the most prominent species predicted for an Enceladus ocean (Zolotov 2007) were identified in E ring and plume grains (Postberg et al. 2009, 2011) by Cassini CDA. These CDA compositional measurements are described in detail in section (3.1.3). The interpretation of the results is shown in figure (15), suggesting that salt rich grains are formed from droplets formed by dispersion from a salty liquid. These droplets would be sustained in the vapor above the liquid and become entrained in the quasi steady flow of gas to the surface through back-pressurized vents.

One possible mechanism for dispersion of droplets from the water could be bubbles of exsolved gases (Matson et al. 2012). The droplets would freeze and be transported to the surface by the nearly steady gas stream through the vents. The salt-poor grains, and especially the nano-grains seen by CAPS (Jones et al. 2009; Hill et al. 2012), could form by homogeneous nucleation from the supersaturated water vapor. In such a scenario the water gas could be formed by evaporation from both the liquid and the surrounding warm ice (Fig. 16). Additional volatile gases could be contributed from the ice, possibly created by clathrate decomposition close to the warm shear-zone and transported by diffusion.

5 Interaction with Saturnian system

5.1 Introduction

Enceladus and its plume do not exist in isolation, but are under the influence of their immediate environments. The moon and plume are constantly bombarded...
by magnetospheric thermal plasma and the energetic particles of Saturn’s radiation belts. This bombardment liberates material through sputtering, affecting the chemistry of the plume. Strong currents, set up between the ionized plume and Saturn itself, provide energy input into the planet’s upper atmosphere. In return, Enceladus is the primary source of material for the E ring, and is also the ultimate source of material contained in the neutral and ionized tori that encircle the planet. These tori exert a strong influence on the nature of the planet’s entire magnetosphere.

Here, we review our current understanding of this complex set of interactions. A schematic overview of the domains directly or indirectly influenced by the moon’s activity is given in Fig. 17.

5.1.1 Discovery of Enceladus’s magnetospheric interaction

Enceladus’s unusual nature as an extremely high albedo object, as revealed by the Voyager probes, made encounters with it a high priority when planning Cassini-Huygens’s tour of the Saturn system. As described below, its coincidence with the densest region of the E ring identified it as the likely source of particles within that ring (Baum et al. 1981). Shemansky et al. (1993) reported the presence of an OH torus, again seemingly associated with Enceladus’s orbit. Evidence of a magnetospheric interaction was not as forthcoming. Pioneer 11 and Voyager 1 and 2 data revealed the presence of depletions in energetic particle fluxes termed micro- and macro-signatures that are also present at other moons’ L shells (van Allen et al. 1980; Carbary et al. 1983).

It fell to Cassini to reveal the extensive nature of this perplexing moon’s interaction with the Saturn magnetosphere. Table 1 lists Cassini’s encounters with Enceladus to date. Encounter E0, in February 2005, was a relatively distant encounter, but during that approach to the moon, the magnetometer (Dougherty et al. 2004) detected a significant deviation in magnetic field strength and direction that indicated that the Enceladus-magnetosphere interaction was a strong one, that the scale of the interaction was larger than the moon itself, and that the moon possessed an ionized atmosphere (Dougherty et al. 2006). As a result, the E2 encounter was lowered, and the data from the magnetometer and other particles and fields instruments began to reveal the true extent of the moon-magnetosphere interaction (Kivelson 2006 and references therein).
5.1.2 Cassini’s encounters with Enceladus

Before providing an overview of Cassini’s results, it should be borne in mind that our exploration of the moon-magnetosphere interaction region is far from complete. As tidally-locked Enceladus resides in a continuous flow of magnetospheric plasma almost comoving with Saturn’s rotation, one hemisphere is constantly bombarded by thermal plasma. As the plume source surrounds the moon’s southern pole, a north-south asymmetry in the distribution of neutral and ionized gas, as well as particulate matter, also exists. Cassini’s trajectory during each encounter with the moon is essentially a straight line on the scale of the interaction region, therefore to fully understand the interaction, many encounters are required. As well as only having limited coverage in terms of encounter geometries, i.e. up- and downstream, north and south, of the moon, as the spacecraft is three-axis stabilized with most instruments having limited fields of view which have tight pointing constraints, we have not yet obtained all observations required to put together a complete picture of the interactions occurring at the moon. In no particular order, the instruments whose results constitute the majority of the material presented here are the Magnetospheric Imaging Instrument, MIMI (Krimigis et al. 2004), the Magnetometer,
Fig. 18 Cassini encounter geometries, separated according to the four types of encounter. Top panels: view looking along the thermal plasma wake, north upwards, corotation flow out of the page, Saturn towards the right. Bottom panels: view from the north; corotation from from top, Saturn towards the right. [The figure pairs need to be boxed to together to make the association more clear]

MAG (Dougherty et al. 2004), the Cassini Plasma Spectrometer, CAPS (Young et al. 2004), the Cosmic Dust Analyzer, CDA (Srama et al. 2004), the Radio and Plasma Wave investigation (Gurnett et al. 2004), the Ion and Neutral Mass Spectrometer, INMS (Waite et al. 2004), and the Ultraviolet Imaging Spectrometer, UVIS (Esposito et al. 2004). For the purposes of investigating the interaction with the magnetosphere, Enceladus encounters can be classified by flyby geometry into four groups, as follows:

- Upstream encounters (E0, E1, E2, E10): Cassini passes upstream of the moon. It does not encounter the thermal plasma wake, and does not pass directly through the plume.
- North-South plume crossings (E3, E4, E5, E6): Cassini’s velocity is primarily north-south in the Enceladus frame. The spacecraft approaches the moon from downstream, then passes through the plume, emerging upstream. E4, E5, and E6 were very similar in geometry.
- Southern encounters (E7, E8, E9, E11): Cassini passes directly through the plume with only a small velocity component in the north-south direction. It remains south of the moon throughout the encounter.
- Northern encounters (E12, E13): Cassini passes directly north of Enceladus with only a small north-south component of motion.
<table>
<thead>
<tr>
<th>Encounter</th>
<th>Neutral source rate</th>
<th>Ion source rate</th>
</tr>
</thead>
<tbody>
<tr>
<td>E0</td>
<td>1600 kg/s</td>
<td>0.2 kg/s</td>
</tr>
<tr>
<td></td>
<td>(Saur et al. 2008)</td>
<td></td>
</tr>
<tr>
<td>E1</td>
<td>200 kg/s</td>
<td>100 kg/s over 30 RE pickup region</td>
</tr>
<tr>
<td></td>
<td>(Saur et al. 2008)</td>
<td>(Pontius and Hill 2006)</td>
</tr>
<tr>
<td>E2</td>
<td>18 (or 36 or 72) kg/s</td>
<td>0.2 kg/s</td>
</tr>
<tr>
<td></td>
<td>(Smith et al. 2010)</td>
<td>(Saur et al. 2008)</td>
</tr>
<tr>
<td></td>
<td>200 kg/s</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(Saur et al. 2008)</td>
<td>3 kgs</td>
</tr>
<tr>
<td></td>
<td>200 kg/s</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(Burger et al. 2007)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>E2 6kg/s</td>
<td></td>
</tr>
<tr>
<td>E3</td>
<td>370 kg/s</td>
<td>1-3 kg/s over 5 RE pickup region</td>
</tr>
<tr>
<td></td>
<td>(Dong and Hill 2009)</td>
<td>(Khurana et al.)</td>
</tr>
<tr>
<td>E4</td>
<td>750 kg/s</td>
<td>1-3 kg/s over 5 RE pickup region</td>
</tr>
<tr>
<td></td>
<td>(Smith et al. 2010)</td>
<td>(Khurana et al.)</td>
</tr>
<tr>
<td></td>
<td>700 kg/s</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(Dong and Hill 2009)</td>
<td></td>
</tr>
<tr>
<td>E5</td>
<td>500 kg/s</td>
<td></td>
</tr>
<tr>
<td>E6</td>
<td>6.5 kg/s lost from neutral torus</td>
<td></td>
</tr>
<tr>
<td></td>
<td>(Leisner et al. 2006)</td>
<td></td>
</tr>
<tr>
<td>E7</td>
<td>6.5 kg/s lost from neutral torus</td>
<td></td>
</tr>
<tr>
<td>N/A</td>
<td>6.5 kg/s lost from neutral torus</td>
<td></td>
</tr>
</tbody>
</table>

Table 3 Estimates of neutral and ionized gas production rates at Enceladus.

5.2 The Plume

5.2.1 Production rate and composition

The processes leading to the generation of the Enceladus plume are described elsewhere in this publication. By injecting significant quantities of neutral gases and solid icy grains into Saturn's magnetosphere, Enceladus acts as a major controller of magnetospheric composition and dynamics (Kivelson 2006). The physical and chemical processes occurring in the plume are numerous.

Estimates of the production rate of neutrals are varied, but most derived values are of the order of hundreds of kg per second. The variability in the quantitative estimates may largely be due to a real variation in the plume activity level; a summary of production rate estimates is provided in Table 3.

5.2.2 Ionization in the Torus and the Plume

Several potential sources of plasma exist in the Enceladus torus and/or in the Enceladus plume, including:

1. photoionization of neutrals by solar extreme ultraviolet (EUV) and x-ray radiation
2. impact ionization of neutrals by energetic magnetospheric ions and electrons
3. photoionization and impact ionization of E-ring grains and satellite (i.e., Enceladus) surfaces.

The main neutral species in the plume, as well as in the dense, main neutral torus located at radial distances close to 4 R_S, is water (H_2O), but water dissociation products (i.e., OH, O, H) are also present (Waite et al. 2006, 2009, 2011; Teolis et al. 2010), particularly in the more extended outer torus (Johnson et al. 2006; Perry et al. 2010; Jurac 2001; Richardson et al. 1998; Melin et al. 2009). Many other minor neutral species were detected in the plume by the INMS (Waite et al. 2009) including CO_2, C_2H_2.(See section xxxx of this review). The primary
ionization reaction can be represented by:

\[ h\nu + H_2O \rightarrow H_2O^+ + e^- \]
\[ OH^+ + H + e^- \]
\[ O^+ + OH + e^- \ldots \]

(5)

The photoionization frequency due to absorption of solar radiation by water, is very close to a value of \( 9 \times 10^{-9} \text{ s}^{-1} \) at the heliocentric distance of Saturn and for solar minimum conditions (c.f., Burger et al. (2007); Cravens, T. E., Ozak, M. S., Campbell, I. P., Perry, M., Rymer A. M. (2011)). Secondary ionization by photoelectrons contributes another 10%, giving a total photoionization frequency of \( 10^{-8} \text{ s}^{-1} \). The ion production rate in the torus and/or plume for low opacity conditions (which is almost everywhere outside a distance of 1 R_S) is this ionization frequency multiplied by the neutral number density. In the core neutral torus, photoionization appears to be the dominant ionization mechanism. Most of the suprathermal electron distribution measured by CAPS ELS (Schippers, P., Andre, N., Johnson, R. E., Blanc, M., (Dandouras), I., Coates, A. J., Krimigis, S. M., Young, D. T. (2009) appears to be explainable by photoelectrons associated with photoionization, at least for electron energies less than about 70 eV. Measured electron distributions and model distributions agree well for \( r \sim 4-5 \text{ R}_S \). The distribution appears to consist of a Maxwellian core (\( n_e \sim 80 \text{ cm}^{-3} \) and \( kT_e \sim 2 \text{ eV} \)), a photoelectron region (20 - 70 eV), and a more energetic tail region (\( E > 70 \text{ eV} \)).

5.2.3 Charged dust grains

During close Enceladus encounters E3 (March 12, 2008), E5 (October 9, 2008), E7 (November 2, 2009), E10 (May 18, 2010), and E13 (December 21, 2011), the CAPS sensors’ apertures were oriented to encompass the ram direction, allowing direct sampling of plume material moving slowly relative to Enceladus. During the first of these encounters, IMS, ELS, and IBS unexpectedly recorded extremely high \( \geq 1 \text{ keV/q} \) count rates within the plume. The IMS and ELS observations were reported upon by Jones et al (2009). Populations of high energy magnetospheric ions and electrons were quickly ruled out as they were only detected by CAPS anodes nearest the local ram direction.

The signatures were therefore almost certainly due to massive ions in the Enceladus plume. To estimate their masses, as is the case at Titan, as ions enter CAPS sensors at known velocities, their measured kinetic energy per charge can be converted to a mass-to-charge ratio (M/q). The E3 encounter, with its encounter speed of 14.4 km s\(^{-1}\), turned out to have recorded negative particles with M/q values of up to 26 600 u/q, and up to 46 500 u/q for positive grains, implying that these particles are charged nanodust. As it is the relative speed of spacecraft and plume material that determines the conversion factor between E/q and M/q, lower speeds allow a greater M/q range to be covered by each CAPS sensor, at the expense of M/q resolution. The encounters with suitable CAPS pointing involved near-constant encounter speeds ranging from 6.2 to 17.7 km s\(^{-1}\). Encounter E10, with its 6.2 km s\(^{-1}\) speed potentially allowed the detection of negative grains with masses up to 144 600 u/q, compared to the fastest encounter E5, where its 17.7 km s\(^{-1}\) speed allowed only masses per charge of up to only 17 700 u/q to be detected.
Data from the E3 and E5 encounters remain the most striking obtained to date, showing very high fluxes up to the top of the sensors' E/q ranges (Jones et al. 2009); at lower energies, low mass positive and negative ions originating in the plume gas itself were also detected (Tokar et al. 2009; Coates et al. 2010a). The ≥ 500 eV/e₀ signatures during these encounters imply that ELS detected negative ions of M/q from 400 up to 26 600 and 17 600 u/q, respectively, and IMS detected positive ions up to 46 500 and 30 700 u/q. The grains’ charge state is undetermined, but if they are singly-charged, they measure up to > 2 nm radius, i.e., orders of magnitude smaller than plume grains detected by other instrumentation (Spahn et al. 2006b). The derivation of the number density of the nanodust is currently ongoing; a direct calculation of this parameter from the raw data is not possible due to the many instrumental effects that need to be taken into account, including, for example, the effects of ELS microchannel plate detection efficiencies for high mass particles.

The source of these grains is the condensation of water vapour that emerges at Enceladus’s south polar region at surface fractures. Modelling of this process (Schmidt et al. 2008) predicts the formation of grains on scales from nm to tens of mm, with the smallest attaining velocities close to the gas speed.

Although some differences between signatures observed by the different elements of CAPS are due to the sensors’ time resolution, there is a clear mismatch between the timing of fine scale structures observed in the positive and negative grain data. In E3 IMS data, one clear, relatively broad flux peak is observed, preceded by a minor isolated peak. In ELS, two main peaks are seen: one detected over ELS’s entire energy range and all anodes, possibly resulting from grain fragmentation inside ELS itself, and a second, more complex signature where the peak energy shifts over time, indicating negative grains’ spatial separation.

The temporal, and therefore spatial differences between positive and negative fine-scale structures are believed to be due to the effects of electromagnetic forces on the nanodust. Initially, the motional electric field associated with the flow of magnetospheric plasma past Enceladus will accelerate negative grains towards Saturn, and positive grains away from the planet. As reported by Jones et al. (2009), at a qualitative level, this appears to agree with the jet source locations catalogued by Spitale and Porco (2007). Detailed modeling of the processes moulding the shape of the jets is ongoing (Arridge et al. 2010), but it is clear that the electromagnetic fields present in the vicinity of Enceladus essentially allow the plume region to act as a mass spectrometer, splitting the particles by mass and charge state. Jets that are initially collimated flows of particles are likely to be splayed into sheets of grains. When Cassini crosses one of these highly modified jets, it may only encounter particles with a very limited range of M/q at a given location, as may have occurred during encounter E10.

One puzzle arising from the CAPS observations is the colocation of positively and negatively charged nanometer particles. The overriding charging process near Enceladus is caused by the plume material’s immersion in Saturn’s corotational magnetospheric plasma, resulting in negative potentials (Kempf et al. 2006). The co-existence of both populations at first appears anomalous. At nanometre scales, however, the charging process can be stochastic, depending on local plasma parameters. In addition, during their subsurface formation in a collisional environment (Schmidt et al. 2008), grains could undergo triboelectric charging; those that condense within the vent even when of the same composition, can acquire opposite charges. Smaller particles tend to charge negative, and larger ones positive (e.g. Duff and Lacks (2008)). Most entrained nanodust particles are likely to have been accelerated to near-gas velocities.

Overall, during both E3 and E5 encounters, CAPS detected negative particles to much lower kinetic energies than for positive particles; if a proxy for lower masses, this supports the picture of triboelectric charging occurring within vents. Although sunlight will have little effect in the near-Enceladus environment, particles’ charge state changes could vary once exposed to the plume and the corotational plasma flow, where plasma parameters can differ significantly. This may have implications for the behaviour of particles under the influence of electromagnetic
forces, as their charge state changes, possibly reversing polarity, grains’ paths can be affected significantly. Grains initially charged positive and deflected away from Saturn may charge negative, causing them to then be accelerated in the opposite direction. The positively-charged population that can be accelerated outwards by the magnetospheric corotational electric field very likely forms a component of the dust streams observed outside Saturn’s magnetosphere (Kempf et al. 2005). 

Farrell et al. (2010) argued that the charge states observed by IMS and ELS may not reflect the initial triboelectric charge state of the detected particles, due to the likelihood of charge modification between their emergence at the surface and their detection at Cassini. Shafiq et al. (2011) also address the issue of grain charging of particles exposed to the plume environment, based on RPWS Langmuir Probe data. The susceptibility of low mass charged grains to electromagnetic forces means that the fine structures in their signatures is expected to differ significantly from those observed by CDA and RPWS, which detect high mass grains.

The presence of the dusty plasma envelope itself is thought to have a significant effect on Saturnian magnetospheric plasma flow. Saturn’s magnetospheric plasma is observed to slow down and be deflected in the vicinity of Enceladus (Tokar et al. 2006). This is at least partially due to the mass-loading of the plasma by fresh pickup ions created in and around the plume, but the presence of the charged grains may also make a significant contribution to this deceleration and deflection of the flow (Farrell et al. 2010). The charged grains’ presence may have further implications: Simon et al. (2011) suggest that the nature of the electrodynamic coupling between Enceladus and Saturn (Pryor et al. 2011) is significantly affected by the grains’ presence. Rather than being an observational curiosity, it appears that this population of particles spanning the mass range between heavy molecules and micron-scale dust has important effects on several processes occurring in Saturn’s inner magnetosphere.

5.3 Electrodynamic coupling

Satellites such as Enceladus that orbit within their parent planets magnetosphere can interact electrodynamically with their surroundings, largely due to the relative motion of the magnetospheric plasma and the moon. One of the key characteristics of such interactions is a perturbation of the planetary magnetic field. Elements of this interaction include the Pedersen and Hall currents generated by ion-neutral collisions, and a pickup current generated by the incorporation of freshly-created ions by the magnetospheric flow. These current are driven by the electric field that drives the plasma \( E \times B \) drift, and are perpendicular to the magnetic field. Closure of these currents is achieved by currents flowing along the magnetic field, which couples the interaction region in the vicinity of the moon to the ionosphere of the planet. The moon-planet current system can be modelled as a standing Alfvén wave propagating along the magnetic field to the planets ionosphere (Neubauer 1980). The locations at which these currents reach the planets ionosphere can be observed as bright auroral footprints at Jupiter from radio to UV wavelengths, where three of the Galilean satellites have a strong enough interaction for the precipitating currents to be observed remotely (Clarke et al. 2002). Enceladus, with its substantial plume and associated ionosphere, clearly has the ingredients for a moon-magnetosphere interaction potentially strong enough to form remotely-observable auroral spots at its two magnetic footpoints which may provide significant local energy input at the top of Saturn’s atmosphere. A search for such a signature in HST UV images obtained between February 2005 and January 2007 was unsuccessful (Wannawichian et al. 2008).

However, Pryor et al. (2011) reported the detection of an auroral spot associated with Enceladus in Saturn’s northern hemisphere. The search for this feature in UVIS data was spurred on by the first detection of field-aligned electrons near Enceladus, during the E4 encounter, when CAPS-ELS observed energetic (1 ∼ 20 keV) field-aligned electrons arriving at Enceladus from the direction of Saturn’s northern hemisphere, together with coincident energetic ions (55-90 keV) being
Enceladus as an active body

observed in the same direction by MIMI. The beams were observed downstream of Enceladus, from > 23.3 to 3.6 Re, where they stopped abruptly despite the viewing geometry still being conducive to field-aligned observations. The beam electrons flickered in energy between populations peaking at ~10 eV and 1 keV; bimodal populations were also briefly observed. These changes appear to be associated with changes in the magnetic field, and hence represent a true change in the field-aligned current density. Pryor et al. (2011) estimated that the Alfvén wave travel time between Enceladus and Saturn’s ionosphere would be 150 s, yielding a distance for the beams to begin of ~3.6 Re, consistent with the observations.

The auroral signature observed by UVIS was at 64.1 ± 0.4 degrees, near to the predicted 64.5 degrees. The northern signature was not, however, always detectable, suggesting a significant variation in brightness, and its southern hemisphere counterpart was not observed at all. When observable, the northern spot varied in brightness by a factor of ~3. The spots size indicated an interaction region extending to ~20 Re downstream of the moon, and up to 20 Re in radial width. The power transferred in this interaction is ~3 · 10^8 W, several orders of magnitude lower than the equivalent interactions at Io and Europa, at ~1012 and ~1011 W, respectively.

Further in situ observations of elements of the electrodynamic interaction region were presented by Gurnett et al. (2011), who reported the detection of Whistler-mode auroral hiss emissions generated by magnetic field-aligned electron beams from a few tens of eV to several keV in energy. Auroral hiss has a characteristic V-shaped feature when observed in frequency-time spectra obtained during moon encounters, and was present in data from 6 of the 12 flybys studied. The basic electrodynamic interaction at Enceladus is expected to consist of planet to moon currents flowing on the Saturn-facing sides of the flux tube, and moon to planet currents running on the anti-Saturn side. Saur et al. (2007) noted a likely important complication of this scenario at Enceladus, as the plume ionosphere to the south of the moon is largely blocked from a direct connection to the northern hemisphere due to the presence of Enceladus itself. They predicted the presence of additional surface currents running between Saturns northern and southern hemispheres on the planet-facing side and in the reverse direction on the opposite side of the interaction region.

Simon et al. (2011) reported that the presence of charged dust in the vicinity of Enceladus could significantly change the nature of the interaction, including a reversal in the sign of the Hall conductivity. The same authors also reported the detection in magnetometer data of the abovementioned blockage of the Alfvén wing by Enceladus itself.

5.3.1 Observations, models and particular aspects of Enceladus plasma interaction

The plasma just upstream of Enceladus, but not in the plume itself (e.g., a radial distance from Enceladus of some tens of Re), as seen during the E2 flyby (Tokar et al. 2006) was observed by the CAPS IMS to consist mainly water group ions (mostly \( \mathrm{H_2O}^+ \), \( \mathrm{OH}^+ \), and \( \mathrm{O}^+ \) with some \( \mathrm{H_3O}^+ \). The measured drift speed is what one would expect for corotation \( \mu_p \sim 26 \text{ km s}^{-1} \) in the Enceladus/Keplerian frame of reference). The observed distribution function was torus-like (or ring-like in velocity space) but with a rather large spread of velocities in the ring (Tokar et al. 2006). Ions created by ionization of neutrals, or by charge exchange reactions with neutrals, are picked up by the magnetic and motional electric fields and form a ring distribution function. The total observed plasma density upstream of Enceladus is about 100 cm\(^{-3}\) (Persoon et al. 2009a; Tokar et al. 2006).

Water group ions were observed in the plume itself by the CAPS IMS, and the INMS experiments (Tokar et al. 2006, 2009; Cravens et al. 2009). The total ion density appears to increase in the plume itself and the flow speed of the ions relative to Enceladus was seen to becomes quite small (Tokar et al. 2009). Figure 19 shows ion count rate versus energy measured by CAPS IMS during the E3 encounter. The ions move slowly with respect to Enceladus and are cold. An approximate ion composition was obtained from CAPS data in this stagnation re-

-3

Enceladus is about 100 cm
Fig. 19 Cassini CAPS IMS count rate versus energy for ions during the close encounters of E3 (top) and E5 (bottom). When the plasma is cold, the energy spectrum is approximately a mass spectrum. Water group ions are evident in the plume. From (Tokar et al. 2009).

\[
\text{H}_3\text{O}^+ + \text{H}_2\text{O} \rightarrow \text{H}_3\text{O}^+ + \text{OH}
\]

\[
\text{H}_3\text{O}^+ \text{ has also observed to be the major species in the inner coma of active comets due to this reaction (Balsiger et al. 1986). The CAPS IMS and the INMS also detected water cluster ions including } \text{H}_3\text{O}^+ - \text{H}_2\text{O and } \text{H}_2\text{O}^+ - \text{H}_2\text{O (Cravens et al. 2009). CAPS ELS measurements have shown that negatively charged, as well as positively charged, ions are present in the plume (Coates et al. 2010b). The range of ion masses is very large and extends up to nanometer sizes (i.e., grains) (Jones et al. 2009). These negative ions are poorly understood. The CAPS ELS also measured the electron flux versus energy in the plume, as shown in Figure 21. The electron distribution is dominated by electrons with energies of about 10-20 eV. Ozak et al. (2012) modeled the electron energy distribution in the plume using a two-stream transport code that included the photoelectrons produced by photoionization of neutrals by solar radiation, and elastic and inelastic (i.e., ionization collisions and electronic, vibrational and rotational excitation collisions). The model-data comparisons (Figure 21) showed that most the electrons in the plume are photoelectrons that are either locally produced or come from Saturn’s magnetosphere outside the plume. The ledge in the CAPS ELS spectrum near 25 eV is due to photoionization of water by solar HeII resonant line photons and such peaks are characteristic of photoelectrons in other space environments (see Coates paper on Titan.).}
Fig. 20 The ion composition measured by the Cassini Ion and Neutral Mass Spectrometer in the plume during the E3 encounter. Mass 19 ions (H3O+) strongly dominate the composition. From Cravens et al (2009).

Fig. 21 The electron energy flux versus energy in the plume during the E3 encounter by the CAPS ELS (blackline with dots). The spectrum modeled using a two-stream transport method is shown as the yellow and red lines (no external electron inputs) or as the green and blue lines (external electron inputs). The purple dashed line shows a Maxwellian distribution. From Ozak et al. (2012).

The external plasma flow has been observed to slow down as it enters the plume, as discussed above. This is expected due to the mass loading of the flow by ionization and charge exchange, which create pick-up ions. Numerical models have been used to explain and interpret this interaction.

5.4 E ring

There is an intimate connection between Enceladus and Saturn’s diffuse E ring, which envelopes the ice moons Mimas (3.07 RS), Enceladus (3.95 RS), Tethys (4.88 RS),
Dione (6.25 Rₚ), Rhea (8.73 Rₚ), and Titan (Rₚ). The ring is composed of predominantly water ice grains (Hillier et al 2007) smaller than 10 μm (Nicholson et al 1996; Kempf et al 2006). Numerical studies of the ring particle dynamics suggest particle lifetimes of less than 200 years (Jurac 2001; Juhász et al 2007) or collisional ejecta originating from the ring moons Almathea and Thebe, whose orbital extremes coincide with the outer boundaries of the rings could be explained by collisional ejecta originating from the ring moons Almathea and Thebe, whose orbital extremes coincide with the outer boundaries of the rings (Feibelman 1967a). Enceladus was proposed to be the dominant source of ring particles because the ring’s edge-on brightness profile has its maximum near the moon’s orbital distance (Baum et al 1981). In infrared edge-on images taken during Saturn’s ring plane crossing (August 1995), de Pater et al (2004) found small density enhancements around the orbits of Tethys and Dione, suggesting that all ring moons may contribute to the ring particle population. Surprisingly, the de Pater et al (2004) finding has not been confirmed by any of the Cassini instruments (Horányi et al 2009). Only in situ data obtained by RPWS and CDA during a close flyby of Rhea may provide evidence that a moon other than Enceladus injects a small amount of fresh dust into the ring (Jones et al 2008). The current view is that Enceladus is by far the strongest, and probably the only, source of E ring particles.

Two dust production mechanisms have been postulated for Enceladus. Interestingly, geyser-like processes were proposed early (Haff et al 1983; Pang et al 1984), although in later years most attention was paid to meteoroidal impact ejection (e.g. Hamilton and Burns 1997 or Juhász and Horányi 2002). This change of focus was probably due to the fact that the structure of Jupiter’s diffuse gossamer rings could be explained by collisional ejecta originating from the ring moons Almathea and Thebe, whose orbital extremes coincide with the outer boundaries of the rings (Burns et al 1999). However, the discovery of the Enceladian dust jets dramatically changed the picture, with a lot of the missing puzzle pieces falling into place. In particular, two observations have intrigued scientists from early on. Firstly, the E ring has an unusual blue colour, implying a narrow size distribution centred around 1 μm (Nicholson et al 1996), while rings resupplied with collisional ejecta would be expected to have a broad size distribution. Secondly, the ring’s vertical scale height is not bound by the orbital extremes of Enceladus but increases with the radial distance to the moon (Showalter et al 1991b). This implies that either the grains’ ejection speed is much larger than the satellite’s escape speed of 207 m s⁻¹, contradicting laboratory data for impact ejecta (Söfler et al 1975; Hartmann 1985), or the inclinations of the ring particles have to grow rapidly after their escape into the ring. Although the latter may be induced by the time-dependent out-of-plane component of the solar radiation pressure (Horányi et al 1992), numerical simulations of the ring dynamics do not reproduce the full extent of the observed vertical scale height (e.g. Juhász and Horányi 2002).

The injection of dust grains from localised sources in the south polar terrain of Enceladus explains many aspects of the ring’s peculiar vertical profile. Since the grains’ ejection speed is much slower than the Enceladus orbital speed vₑ, the orbital elements of fresh ring particles and the moon are almost identical. However, because plume particles are ejected roughly orthogonally to the ring plane, the minimum inclination of the ejected grains is \( i_{\text{min}} = \frac{R_{\text{Hill}}}{v_e} = 0.23° \), where \( R_{\text{Hill}} = 948 \text{ km} \) is the Hill radius of Enceladus, characterising the range of moon’s gravitational influence, and \( v_e = 3.95 \text{ Rₚ} \) is the moon’s mean orbital distance from Saturn. Particles ejected faster than the three-body escape speed from Enceladus of \( v_{\text{esc}} = 207 \text{ m s}^{-1} \) may escape to the ring and have an initial orbit inclination of

\[
i \approx \left( 0.23° + \frac{v_e^2 - v_{\text{esc}}^2}{v_e^2} \right)^{1/2}.
\]

Numerical studies by Kempf et al (2010) showed that, in order to escape to the ring, plume particles need be launched faster than 224 m s⁻¹, as slower grains recollide with Enceladus during their first orbit. Furthermore, the initial inclination of grains launched slower than 224 m s⁻¹ will be altered by three-body effects, because these grains traverse the Enceladus Hill sphere during their first ring
plane crossings. By tracing the trajectories of plume particles numerically, Kempf et al. (2010) computed the spatial distribution of the plume particles in the vicinity of Enceladus. The model includes the effects of the higher moments of Saturn’s gravitational field as well the Lorentz force due to Saturn’s rotating magnetic field acting on the charged grains. The time-dependent grain charge due to the ambient plasma and the solar UV radiation is computed simultaneously with the particles’ equations of motion. Finally, the distributions of the initial speeds and grain sizes were as described by the Schmidt et al. (2008) model. The resulting vertical ring profile matches the measurements by the Cassini dust detector CDA well. Figure 22 compares the simulations with CDA data obtained during a steep Cassini traversal through the ring plane at approximately the orbital distance of Enceladus (Kempf et al. 2008). The kinks in the vertical profile are associated with individual Enceladian dust jets and are reproduced well by the simulations.

The vertical ring profile is not the only imprint of the localised dust injection by Enceladus’ south pole plume. The radial brightness profiles derived from earth-bound observations indicated that the densest point of the E ring lies outside the orbit of Enceladus, displaced by 10^4 km (de Pater et al. 2004), while in situ measurements by Cassini CDA found the displacement to be > 3000 km (Kempf et al. 2008). Numerical simulations before the discovery of the Enceladian plume failed to reproduce this observation. Juhász et al. (2007) showed that the initial inclinations of the plume particles decrease the likelihood of the particles re-colliding with Enceladus, allowing them to migrate outward by drag forces exerted by the ambient plasma. The imprint of the Enceladus dust jets is also clearly visible in Cassini images of the E ring (Fig. 23).

Despite the remarkable progress since the discovery of the Enceladus plume there are many phenomena still not understood. One important open question is whether the Enceladus geysers are really the only source of E ring dust. In situ data imply a dust escape rate to the ring of 0.2 kg s^{-1} (Spahn et al. 2006b), which is surely enough to maintain a steady-state E ring. However, it is difficult to understand why the influx of interplanetary meteoroids does not produce noticeable amounts of collisional ejecta. Spahn et al. (2006a) reevaluated the expected ejecta production rates of the E ring moons in the light of the new Cassini data. The authors conclude that impactors of interplanetary origin produce ~ 0.1 kg s^{-1}, while
Fig. 23  Cassini image of the densest region of the E ring taken at a distance of about $2.1 \cdot 10^6$ km from Enceladus (NASA/JPL/SCI PIA 08321). The moon is within a dust torus fed by the Enceladian ice jets. The strands emerging from the moon form the various micro-signatures apparent in the ring’s vertical (see e.g. Fig. 22) and radial profiles recorded by the Cassini in situ instruments such as RPWS (Kurth et al. 2006) or CDA (Kempf et al. 2008). Tethys is visible to the left of Enceladus.

E ring impactors produce $\sim 10$ kg s$^{-1}$ of fresh dust. If this were the case, collisional ejecta would still be the dominating E ring dust source. However, there is only circumstantial evidence for ejecta in the Cassini data. The best fit of the CDA plume density profile of grains $\geq 1.6 \mu m$ obtained during the first close Enceladus flyby in 2005 (Spahn et al. 2006b) to numerical plume models is obtained by assuming an additional ejecta source with a peak rate of 0.5 s$^{-1}$ at closest approach to Enceladus. A similar conclusion was reached by Spahn et al. (2006b), who compared the CDA data with a plume formed from grains emerging from a circular area of $30^\circ$ centred at the moon’s south pole. If there is really an ejecta source of this strength, then most of the ejecta grains must be produced by ring particle collisions. In this case, however, ejecta from the ring moon Tethys should also form a dust torus, which hasn’t yet been observed.

The large number of Cassini traversals through the inner E ring allowed the Cassini in situ instruments, mainly RPWS and CDA, to monitor the time variation of the ring particle density. Kurth et al. (2006) reported on one equatorial and 5 vertical passages through the inner E ring during which RPWS measured ring particle impacts, while Kempf et al. (2008) published CDA dust density data for grains $\geq 0.9 \mu m$ recorded during 2 equatorial and 6 vertical Cassini ring traversals. The data by the latter team provides evidence for a variation in the ring’s peak density of 20% within 37 days. It is not clear whether the observed variations stem from azimuthal variations within the dust torus or from a time-variable dust plume. Nevertheless, the in situ ring data indicate that the Enceladus dust source supplies the ring with fresh dust at a surprisingly constant rate.

5.5 Inner magnetosphere

5.5.1 Energetic particle interaction

Enceladus orbits Saturn within the planets radiation belts (Roussos et al. 2011) and references therein). The moons surface and plume material are therefore continuously bombarded with energetic particles. The interpretation of the energetic particle signatures is complicated by the varying drift rates of charged particles
in Saturn's magnetosphere. All protons and other positive ions, plus low energy electrons, have a net drift motion around Saturn in the same sense as the planet's rotation. In the rest frame of the corotating thermal plasma, electrons drift opposite to the direction of planetary rotation. At energies around 1 MeV, this counterdrifting motion is of a large enough magnitude for electrons to reverse direction in the inertial frame too; electrons of > 1 MeV travel around Saturn in the direction opposite to the planets rotation. In the Enceladus frame, the low-energy thermal plasma is continuously overtaking the moon; the moons trailing hemisphere is therefore exposed to bombardment by low energy ions and electrons. The bulk motion of energetic electrons is however in the opposite sense, so Enceladuss leading hemisphere is most exposed to these particles.

As is seen in the vicinity of the orbits of other large icy moons, the absorption of energetic particles leads to the formation of narrow cavities of reduced energetic particle number densities. Due to the rapid bounce motion of energetic particles (> tens of keV), these cavities extend to higher latitudes, well out of the plane of the moons orbit. On crossing these cavities, instruments on Cassini detect dropouts in energetic particle fluxes; these are seen as a drop in foreground particles by MIMI-LEMMS, but also cause background decreases in certain CAPS and UVIS data. Such cavities that persist over limited longitude ranges and time periods are referred to as microsignatures. First observed by Pioneer 11 and the Voyagers at the L shells of several icy moons (van Allen et al 1980; Carbary et al 1983), these signatures were successfully modelled in detail by (Paranicas and Cheng 1997). On arrival at the Saturn system, such signatures were clearly observed by MIMI-LEMMS on numerous occasions (Paranicas et al 2005; Jones et al 2006).

Due to magnetospheric particle diffusion, the microsignatures gradually refill, and those of Enceladus are usually eroded while being carried around Saturn before returning to the location of the moon. The energetic electron microsignatures that form upstream of Enceladus, at energies greater than ~ 1 MeV, are seen to be relatively sharp and, as they match the diameter of Enceladus in scale, are largely unaffected by the presence of the plume (Jones et al 2006).

Ahead of the moon in the Keplerian sense, or downstream of the moon in terms of the corotation flow, is a wake of low-energy ions and electron. To date, Cassini has passed through this region during encounters E3 to E6 inclusive. In CAPS-ELS data, a clear absence of thermal electrons exists during the wake traversal. At slightly higher energies, i.e. tens of keV or more, where electrons execute significant bounce motion, the wake extends north and south from the moons orbital plane. The profiles of the microsignatures at there energies are generally less sharply-defined than the high energy microsignatures seen upstream of the moon, and exhibit widths in excess of the diameter of Enceladus. It is believed that these intermediate-energy electrons are partially absorbed by the plume. Variability in the depth and width of these signatures is interpreted as evidence for variability in the plume density, and hence in its activity level (Jones et al 2006).

As energetic ions drift around the planet at such a high rate, the rate of absorption by the moon can be comparable to the rate at which new ions are created or drift into the absorption region. Persistent depletions termed macrosignatures are observed in ions of several MeV (Jones et al 2006). The rate at which microsignatures refill, and shift in radial position, can reveal significant information on the magnetosphere (e.g. Roussos et al 2007).

The exposure of Enceladus surface to the energetic particle environment has observable effects. Early works, e.g. Lanzerotti et al 1983, considered the production of a neutral torus encircling Saturn that consists of material sputtered from the surfaces of Enceladus and other icy moons. The sputtering rates resulting from this exposure to energetic particles have been calculated in detail (Johnson et al 2008); it should be noted that this process affects plume and E-ring grains as well as the moons surface ices. The energy spectra that the surface of Enceladus is exposed to was computed by Paranicas et al. (in press), based on observations made by the MIMI instrument. Although not yet studied in detail for Enceladus, we note the possible importance of surface charging (Roussos et al 2010); at Enceladus location in the inner magnetosphere, incident plasma will generally tend to
charge the moons surface negative, overwhelming photoemission that charges the surface to a positive potential.

Enceladus exhibits a subtle albedo asymmetry (Buratti et al. 1990; Verbiscer and Veverka 1994; Schenk et al. 2011), and generally displays stronger water absorption bands on the Keplerian-leading hemisphere (Verbiscer et al. 2006), at least partially due to the irradiation of its surface. Global colour mapping also reveals the presence a significant colour asymmetry in the form of depressed IR/UV ratios, i.e. more bluish terrain, centered on the south pole and extending northwards at two longitude bands at ~40°W and ~220°W. The colour patterns are distinct from neighbouring large icy satellites; no darkening of the trailing hemisphere is observed. It is likely that the possible near-continuous deposition of plume grains on the surface is responsible for this pattern (Kempf et al. 2010), largely masking the effects of irradiation. The irradiation of the surface by magnetospheric energetic particles could transform crystalline surface ice to an amorphous state (Brown et al. 2006).

5.5.2 Neutral and plasma tori

The emissions of Enceladus create the E-ring and a cloud of neutrals that form a torus centered on the orbit of Enceladus. This neutral cloud then supplies most of the ions in Saturn's inner magnetosphere. In this section we discuss the neutrals and low-energy ions in the inner magnetosphere and provide an overview of their transport and loss processes.

The first strong evidence of a neutral cloud was a neutral-water dissociation product, OH, observed from 3 to 8 RS in Hubble Space Telescope (HST) spectral observations. Shemansky et al. (1993) reported these hydroxyl neutral molecules, which indicated the presence of water, in quantities more than predicted by theoretical estimates of sputtering of the rings and icy satellites (Ip 1997; Jurac et al. 2002). Later observations and modeling by Jurac (2001) showed that OH density peaked near 4 RS, suggesting that Enceladus could be a source of H$_2$O for the Saturn system. The OH observations with HST were sparse, and building sufficient statistics required summing data from several observation periods. The initial models of OH and its parent, H$_2$O, identified a radial and vertical (normal to the equatorial plane) structure to the neutrals, but assumed azimuthal symmetry (Richardson et al. 1998; Jurac 2001; Johnson et al. 2006).

Evidence of IM ions predates the discovery of the neutral cloud. During the Voyager missions, the Voyager Plasma Science experiment data showed ions with a mass greater than H$_2$ in Saturn's inner magnetosphere (Lazarus and McNutt 1983). The energy-per-charge measurements were insufficient for unambiguous identification, but the nitrogen atmosphere at Titan suggested that these heavy ions were N$^+$. The neutral OH by Shemansky et al. (1993) indicated that the heavy ions were likely to be water-group ions. Cassini instruments have confirmed this identification with remote UV measurements of neutral O molecules (Esposito et al. 2005), in situ ion measurements of water-group ions (Tokar et al. 2006; Young et al. 2005), the discovery of water vapor emanating in plumes from Enceladus (Dougherty et al. 2006; Hansen et al. 2006; Porco et al. 2006; Waite et al. 2006), and the measurement of H$_2$O$^+$ ions in the plume (Cravens et al. 2009).

Transport and loss When neutral molecules leave the surface of Enceladus, they become individual particles in orbit around Saturn. For some of the larger jets, the gas may be collisional for 10s of meters, but then becomes collisionless. There are several possible fates of these particles:

1. **Collision with Enceladus or some other icy satellite.** Although it is likely that vapor leaves the surface of Enceladus with different velocities (see Chapter 4), most of the vapor will have speeds greater than the escape velocity, about 240 m/s. (Lower velocities will fall directly back to the surface.) However, as described in the E-ring discussion, molecules that initially escape Enceladus may impact Enceladus within a few orbits, usually within an integer multiple of 16.5 hours, half of Enceladus orbital period. After dispersing away from
Enceladus as an active body

Enceladus, particles may be swept up by other moons or by the rings. Indeed, at 180,000 km, the inner boundary of the neutral cloud coincides with the G ring, which is the outermost ring (excluding dust rings).

2. Dissociation. Electrons, ions, photons, and, less frequently, other neutrals, dissociate neutral molecules into smaller molecules or atoms. Sometimes, ionization also causes dissociation, creating a neutral product along with the ion. This fate changes the composition of the neutral cloud but does not reduce its density.

3. Charge exchange with an ion in the magnetosphere

4. Ionization (excluding charge exchange)

5. Ejected from the Saturn system

Several processes contribute to dispersing neutral molecules. Many are associated with the processes already discussed at loss mechanisms.

1. Initial velocity. A molecule with a velocity of 500 m/s above escape velocity will reach 10,000 km above and below the equatorial plane. Molecules with just 100 m/s excess velocity in the along-track direction will drift with respect to Enceladus.

2. Neutral-neutral collisions. Originally considered negligible, Farmer (2009) showed the importance of this process and subsequent modeling confirmed that neutral-neutral collisions contribute at least half of the dispersion of the neutral cloud. The mean free path is large, 104 to 105 km, but there are still several collisions each orbit around Saturn. These collisions can contribute nearly all of the

3. Velocity associated with ionization byproduct.

4. Charge exchange

Cassini observations in 2005 (Dougherty et al. 2006; Hansen et al. 2006; Waite et al. 2006) discovered water emanating from the south pole of Enceladus, verifying Enceladus as the source of the observed OH and predicted H₂O neutrals.

Distribution The distribution of ions and neutrals in Saturn's IM and their distribution within the water group is sensitive to loss and transport assumptions, but there are direct measurements that provide calibration points for the models. Supporting direct observations of O, OH, and H₂O remote from Enceladus include CAPS data (ions) averaged over several orbits and a few INMS observations of neutrals and ions. There are observations limited to a single species, including UVIS measurements of O and the HST measurements of OH (Shemansky et al. 1993). The resolution of measurements from CAPS is barely sufficient to quantify the separate ion species within the water group. Modeling (Jurac 2001; Smith et al. 2010) implies a distribution amongst water-group neutrals. Additional transport mechanisms affect the ions, so their distribution is less well determined by models.

Since source of the neutrals is highly localized, the neutrals are unlikely to be azimuthally symmetric. Recent models are investigating the geometric dependence of the neutrals (Leisner 2009; Smith et al. 2010), but the data available are insufficient to constrain the azimuthal distribution in the models.

The ion composition of the plasma in either the torus or plume can be altered from its primary state (i.e., H₂O⁺, OH⁺) by ion-neutral collisions (i.e., charge exchange and chemical reactions). Prior to the Cassini mission models of the inner magnetosphere (e.g., Scarf, F. L. and Frank, L. A. and Gurnett, D. A. and Lanzerotti, L. J. and Lazarus, A. and Sittler, Jr., E. C. (1984); Lazarus and McNutt 1983; Richardson et al. 1998) predicted the dominance of water group ions for the ion composition. More recently, Fleshman et al. (2010) and Delamere, P. A. and Bagenal, F. (2008) have modeled the composition and energetics of the Enceladus torus, pointing out the importance of ion-neutral collisions and chemistry. The CAPS IMS experiments demonstrated that water group ions dominate the composition with the most abundant species being H₂O⁺ and O⁺ (Jr. et al. 2006; Wilson et al. 2008; Young et al. 2005; Thomsen, M. F., Reisenfeld, D. B., DeLapp, D. M., Tokar, R. L., Young, D. T., Crary, F. J., Sittler, E. C., McGraw,
M. A. Williams, J. D. (2010)). H$_3$O$^+$ ions are present in the torus, but at relatively lower densities compared with the plume plasma (Wilson et al. [2008]; Tokar et al. [2006]).

The ion distribution function observed in the Enceladus torus appear to be pancake-like in velocity space (Wilson et al. [2008]; Tokar et al. [2006]). The ions drift at roughly the co-rotation speed (\(\sim 45 \text{ km s}^{-1}\)), e.g., Figure 24. The perpendicular average thermal velocity is about 20-30 km s$^{-1}$ with the parallel velocities being less than perpendicular velocities. Ions are initially created at rest with respect to the neutrals (moving at close to the Keplerian speed) but then they are picked up by the magnetospheric motional electric field and the magnetic field forming a ring distribution. But given that a simple ring distribution in velocity space is not observed, the ion distribution functions must evolve (or partially thermalize) due to Coulomb collisions and/or wave-particle interactions. Note that charge exchange collisions remove ions but replace them with ions that are at rest in the neutral frame of reference. This has the effect of removing momentum from the ion population and increasing the ion thermal energy. The electron energy distribution measured in the Enceladus torus, like that in the plume, appears to be predominantly consist of few eV electrons (Sittler, E. C., Ogilvie, K. W., Scudder, J. D. [1983]; Young et al. [2005]; Rymer, A. M., Mauk, B. H., Hill, T. W., Pararicas, C., Andre, N., Sittler, E. C., Mitchell, D. G., Smith, H. T., Johnson, R. E., Coates, A. J., Young, D. T., Bolton, S. J., Thomsen, M. F., Dougherty, M. K. [2007]; Schippers, P., Andre, N., Johnson, R. E., Blanc, M., (Dandouras), I., Coates, A. J., Krimigis, S. M., Young, D. T. [2009]; Delamere, P. A. and Bagenal, E. [2008]). (Schippers, P., Andre, N., Johnson, R. E., Blanc, M., (Dandouras), I., Coates, A. J., Krimigis, S. M., Young, D. T. 2009) fit a typical CAPS ELS electron spectrum (at \(r \sim 5 R_S\)).

1. a cold approximately Maxwellian core
2. a photoelectron component (explaining the observed ledge near an energy of 25 eV - as discussed earlier, such a feature is also present in the plume electron spectrum)
3. a more energetic tail with energies in excess of 50-100 eV.

The fluxes for the energetic tail were quite small however. RPWS Langmuir probe data demonstrated that the core population has a temperature of \(T_e \sim 1-3\text{ eV near 4} R_S\) (Gustafson, G., Wahlund, J. E. [2010]. Cravens, T. E., Ozak, M. S., Campbell, I. P., Perry, M., Rymer A. M. [2011] numerically modeled both the suprathermal electrons and the colder core electrons. The suprathermal electron flux in the torus appears to be very similar to that in the plume (i.e., Figure 21). The heating of thermal electrons is mostly due Coulomb collisions with hot pick-up ions (Sittler, E. C., Ogilvie, K. W., Scudder, J. D. 1983) Gustafson, G., Wahlund, J. E. [2010] but Coulomb collisions with photoelectrons contributes about 20% (Cravens, T. E., Ozak, M. S., Campbell, I. P., Perry, M., Rymer A. M. 2011). The core electrons are cooled mainly by inelastic collisions with water (i.e., rotational and vibrational excitation) (Gustafson, G., Wahlund, J. E. 2010 Cravens, T. E., Ozak, M. S., Campbell, I. P., Perry, M., Rymer A. M. 2011). Cravens, T. E., Ozak, A. 1986).

5.5.3 Low energy electron enhancements

Low energy electron features, termed spikes, have been observed as discrete, short duration enhancements in the low energy plasma density. Their locations suggest an association with Enceladus L shell and have been observed over a wide range of latitudes. The spikes reach over 100 eV in energy and length scales up to hundreds of kilometres. They have been observed continually since 2005 so do not appear to be seasonal. There are no obvious enhancements parallel to the magnetic field direction, signifying that that the electrons responsible for the spikes are not necessarily field aligned. The spikes appear to be more prevalent outside Enceladus L shell and tend to be more common near the moon in longitude. Some of these spikes can be resolved into dispersed features when viewed in data that has had
the penetrating radiation removed. The spikes have varying characteristics; some show energy dispersion, sometimes the peak energy differs, some are single blobs whereas others are clusters of discrete spikes. This suggests that several formation mechanisms may cause these signatures. The fact that spikes cluster around the moons L shell strongly suggests that at least some of them are linked to Enceladus.

The process(es) by which these features are formed, and the cause for their creation, are unidentified at present. However, there are several under consideration. Spacecraft potential changes, hence spacecraft charging, due to energetic flux variations associated with microsignatures [Khurana et al. 1987] have been ruled out, as the spikes occur before and after microsignatures.

The surface of Enceladus is very likely charged; the surface charging of Saturnian icy moons has been addressed by Roussos et al. [2010], and is expected to have a negative potential due to the high density of the plasma impinging upon it. The degree of negative charging varies across the surface, and changes as the moon orbits Saturn due to the direction of incident sunlight changing [Roussos et al. 2010]. Electrons flowing along magnetic field lines connected to the moon encounter this negative potential and, if their energy is less than the magnitude of

![Image](image-url)
the negative potential, are reflected. Secondary electrons are also formed from the acceleration due to the electric field near the moon’s surface. The electron beam would have an energy corresponding to the moon’s surface potential and electrons would only be reflected when their energies are less than the surface potential. Surface charging has been detected at another kronian moon, Rhea (Jones et al., in preparation).

The low energy electron enhancements may result from electron reflection from Enceladus’ surface or even at the plume itself. Negatively charged plume grains may collectively reflect incident low energy electrons. However, the potential levels implied by the spikes peak energies do seem higher than the levels expected; Enceladus’ plume is expected to charge negatively only to about -2 V.

Spikes have also been seen in relation to injection events (DeJong et al. 2010). Another possible cause for the spikes may be related to the moon and plumes absorption of a portion of the local plasma population. This may indicate a plasma instability that causes a small-scale injection event; i.e. the moon itself or the presence of its wake may be responsible for the formation of these dispersion features. Either the dispersion features are created at the moon L shell or the plasma instability causes a flux tube to move inwards without the corresponding outward movement of another flux tube. One limitation of looking for these dispersion events at 4RS is that identification becomes challenging due to high energetic electron background levels as we travel closer and closer to the planet Saturn.

Finally, some of the spikes could be fossil signatures of the moon-planet electrodynamic interaction. Supporting this is the fact that a few of the electron features are seen to extend up to high energies, more than 3 keV, and are dispersed in energy, indicating an age of several hours or more. The electrodynamic interaction is known to accelerate electrons (Pryor et al. 2011), but once the interaction ceases to operate, the accelerated electron population may drift near Enceladus’ L shell where it remains observable for several hours or more.

5.6 Outer magnetosphere and solar wind

Enceladus plays a major role in controlling global magnetospheric dynamics at Saturn. Contrary to some expectations prior to Cassini’s arrival at Saturn, it appears that Titan’s role as a source of neutrals and ions in planets magnetosphere is relatively small: few N+ ions are observed. The main ring system too is a negligible source of plasma, as the population of O2+ expected from a ring source is weak. Finally, the appears to be little input of material from the solar wind in the form of alpha particles, H2+ and H3+. This leaves Enceladus as the primary source of magnetospheric plasma, both directly, from the ionization of plume material in the moons immediate vicinity, and indirectly, through the ionization of the neutral torus that results from Enceladus expulsion of neutral gas and material liberated from E ring grains. The neutral torus has its peak number density at the orbit of Enceladus; these neutrals spread radially inwards and outwards, as evidenced by the OH profile observed by HST (Johnson et al. 2006). In situ CAPS observations show that water ions are the dominant ion species out to 20 Saturn radii from the planet.

The distribution of magnetospheric plasma is governed by equilibrium diffusion (Persoon et al. 2009b). In the model of Persoon et al. (2009b), which fits the observations well, the effect of gravity is negligible, and the diffusion is dominated for water group ions by the centrifugal force. The ambipolar electric field is significant for protons <7 RS from the planet, and magnetic mirroring is dominant for the same hydrogen ions at greater than 7 RS. The model predicts that cold water group ions are confined to low latitudes by the centrifugal force. The addition of water group ions to the rotating magnetosphere leads to the deceleration of the flow at the radial distance of Enceladus (Wilson et al. 2009).

The mass-loading of the inner and middle magnetosphere leads to this region being unstable to the radial interchange of magnetic flux tubes. Magnetic flux tubes carrying cold, heavy ions migrate outwards to be replace by hotter, low
density plasma moving towards Saturn (e.g. Hill et al. (2005); Burch et al. (2005)); this process has been modelled (Hill 2008).

5.6.1 Oxygen at Titan

We note that material originating at Enceladus may have significant effects on the atmospheric chemistry occurring at Titan: Sittler et al. (2009) proposed that, as described above, the dissociation and ionization of water vapour in the plume leads to the formation and ionization of oxygen in the magnetosphere. The flux of O\(^+\) at Titan is estimated to be \(\sim 1 \times 10^6\) ion cm\(^{-2}\) s\(^{-1}\) (Hartle et al. 2006). The vast majority of this O\(^+\) is believed to go towards the formation of CO or CO\(_2\)^+ in Titans atmosphere, but 0.1% can be incorporated into cage-like molecules such as C\(_{60}\) fullerenes. Heavy aerosol particles fall through the atmosphere, forming haze layers, and they eventually reach the surface, where the oxygen is available to partake in chemical reactions. Sittler and colleagues suggest that the presence of oxygen atoms on Titan, originating at Enceladus, could therefore provide the basis for pre-biological chemistry.
References


Canup RM (2010) Origin of Saturn’s rings and inner moons by mass removal from a lost
Carbary JF, Krimagis SM, Ip WH (1983) Energetic particle microsignatures of Saturn’s satel-
Charnoz S, Salmon J, Crida A (2010) The recent formation of Saturn’s moonlets from viscous
spreading of the main rings. Nature465:752–754, DOI 10.1038/nature09096
Charnoz S, Crida A, Castillo-Rogez JC, Lainey V, Dones L, Karatekin O, Tobie G, Mathis S,
Le Poncin-Lafitte C, Salmon J (2011) Accretion of Saturn’s mid-sized moons during the
viscous spreading of young massive rings: Solving the paradox of silicic-poor rings versus
Choukroun M, Kieffer SW, Lu X, Tobie G (2013) Clathrate Hydrates: Implications for Ex-
change Processes in the Outer Solar System, p 409. DOI 10.1007/978-1-4614-3076-6_12
Saturn suprathermal O$_2^+$ and mass-28$^+$ molecular ions: Long-term seasonal and solar
jgra.50383
Clarke JT, Ajello J, Ballester G, Ben Jaffel L, Connerney J, Gérard JC, Gladstone GR, Grodent
D, Pryor W, Trauger, Waite JH (2002) Ultraviolet emissions from the magnetic footprints
Coates AJ, Jones GH, Lewis GR, Wellbrock A, Young DT, Crary FJ, Johnson RE, Cassidy
10.1016/j.icarus.2009.07.013
Coates AJ, Jones GH, Lewis GR, Wellbrock A, Young DT, Crary FJ, Johnson RE, Cassidy
10.1016/j.icarus.2009.07.013
j.icarus.2007.01.010
Chemie der Erde / Geochemistry 68:1–29, DOI 10.1016/j.chemer.2008.01.003
Plume ionosphere of Enceladus as seen by the Cassini ion and neutral mass spectrometer.
Planet Space Sci 34(10):961–970
the Enceladus torus. J Geophys Res 116(A9):A09,205
Icarus73:66–79, DOI 10.1016/0019-1035(88)90085-1
icarus.2009.07.013
Cunningham E (1910) On the Velocity of Steady Fall of Spherical Particles through Fluid
1910.0024
a liquid intergranular phase: rate-controlling processes in the diacolation creep regime.
and G rings during Earth’s ring plane crossing in August 1995. Icarus 172:446–454, DOI
10.1016/j.icarus.2004.07.012
DeJong AD, Burch JL, Goldstein J, Coates AJ, Young DT (2010) Low-energy electrons in Sat-
urn’s inner magnetosphere and their role in interchange injections. Journal of Geophysical
Research (Space Physics) 115:A10229, DOI 10.1029/2010JA015510
Delamere, P A and Bagenal, F (2008) Longitudinal plasma density variations at Saturn caused
by hot electrons. Geophys. Res. Lett.35:3107
Abstracts p B6
Dong Y, Hill TW, Teolis BD, Magee BA, Waite JH (2011) The water vapor plumes of Ence-
Science 307(5713):1266–1270


DOI 10.1007/978-1-4020-9217-6_16

Howett CJA, Spencer JR, Pearl J, Segura M (2011) High heat flow from Enceladus’ south polar region measured using 10–600 cm–1 Cassini/CIRS data... J Geophys Res 116:E03001


Enceladus as an active body


waves, and suprathermal charged particles in Saturn’s inner magnetosphere,

plume variability and the neutral gas densities in Saturn’s magnetosphere. Journal of
Geophysical Research (Space Physics) 115:A10252, DOI 10.1029/2009JA015184

plume variability and the neutral gas densities in Saturn’s magnetosphere. Journal of
Geophysical Research 115(A):10-252

Smith-Konter B, Pappalardo RT (2008) Tidally driven stress accumulation and shear failure

10.1006/icar.1995.0097

to small-scale convection on Earth and other terrestrial planets. J Geophys Res 105:21,795–
21,818, DOI 10.1029/2000JB900197

Sotin C, Parmentier EM (1989) Dynamical consequences of compositional and thermal
10.1029/GL016i008p00835

Spahn F, Albers N, Hörning M, Kempf S, Krivov AV, Makuch M, Schmidt J, Seiß M, Sremčević
M (2006a) E ring dust sources: Implications from Cassini’s dust measurements. Planet
Space Sci 54:1024–1032, DOI 10.1016/j.pss.2006.05.022

Spahn F, Schmidt J, Albers N, Hörning M, Makuch M, Seiß M, Kempf S, Srama R, Dikarev V,
Helfert S, Moragas-Klostermeyer G, Krivov AV, Sremčević M, Tuzzolino AJ, Economou T,
Grün E (2006b) Cassini Dust Measurements at Enceladus and Implications for the Origin


Spencer JR, Barr AC, Esposito LW, Helfenstein P, Ingersoll AP, Jaumann R, McKay CP,
DOI 10.1007/978-1-4020-9217-6_21

Spencer JR, Howett CJA, Verbiscer AJ, Hurford TA, Segura ME, Pearl JC (2011) High-
Resolution Observations of Thermal Emission from the South Pole of Enceladus. In: Lunar
and Planetary Institute Science Conference Abstracts, Lunar and Planetary Institute
Science Conference Abstracts, vol 42, p 2553

Spitale JN, Porco CC (2007) Association of the jets of Enceladus with the warmest regions on

53:319–331

DOI 10.1016/0019-1035(83)90152-5

Srama R, Ahrens T, Altobelli N, Auer S, Bradley J, Burton M, Dikarev V, Economou T,
Fechtig H, Gorlich M, Grande M, Graps A, Grün E, Havnes O, Helfert S, Horányi M,
Igenbergs E, Jessberger E, Johnson TV, Kempf S, Krivov AV, Krüger H, Mocke-Ahreep

Srama R, Ahrens TJ, Altobelli N, Auer S, Bradley JG, Burton M, Dikarev VV, Economou
McDonnell JAM, Möhlmann D, Morfill GE, Müller M, Roy M, Schäfer G, Schlotzhauer G,
DOI 10.1007/s11214-004-1435-5

Stephan DR, Freeman J, May DA (2009) Origin of ice diapirism, true polar wander, subsurface
ocean, and tiger stripes of Enceladus driven by compositional convection. Icarus 202:689–
680, DOI 10.1016/j.icarus.2009.03.017

Stevenson DJ (2008) Constraints on Tidal Heating in Enceladus. AGU Fall Meeting Abstracts
p D4

quartz sand - Distribution and shock metamorphism of ejecta. J Geophys Res 80:4062–
4077, DOI 10.1029/JB080i029p04062

of ice grains and gas distribution in the Enceladus plume by Cassini’s Ion Neutral Mass
2009JA015192

icy satellites and their significance. Icarus 190:573–584, DOI 10.1016/j.icarus.2007.03.012

Thomson, M F, Reisenfeld, D B, Delapp, D M, Tokar, R L, Young, D T, Crary, F J, Sittler,
magnetosphere. J Geophys Res 115(A1020)


