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Satellite Atmospheres

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19.1 INTRODUCTION

Study of the Galilean satellite atmospheres is a field that has blossomed in the 25 years since the last major retrospective of the jovian system (Gehrels 1976). At the time of that book, only one of these satellites, Io, was thought to possess an atmosphere, although a stellar occultation had suggested the possibility of a tenuous atmosphere on Ganymede as well (Carlson et al. 1973).

The discovery of the Io plasma torus (Kupo et al. 1976) and the in situ characterization of the plasma in the jovian magnetosphere by the *Voyager* spacecraft in 1979 made it clear that all of the Galilean satellites are exposed to a harsh radiation environment (see Chapter 23). The collisional impact of magnetospheric ions with these satellites erodes their surfaces and sputters atoms and molecules to create atmospheres and/or exospheres (see Chapter 20). Surface processes such as sublimation and active volcanoes contribute as well.

The discovery and confirmation that all these satellites possess tenuous atmospheres has been accomplished by three of NASA's largest and most productive space missions of the last 25 years: *Voyager*, Hubble Space Telescope (HST) and *Galileo*. Numerous ground-based observations have also contributed to the substantial advances in our understanding of these atmospheres. Whereas in 1975 two species (sodium and potassium) were known to be present on Io, today nine atomic and molecular species have been detected, and 2-4 species have also been detected at each of Europa, Ganymede and Callisto.

We present in this chapter a summary of our current knowledge about the atmospheres of Io, Europa, Ganymede, and Callisto, with a focus on the discoveries and progress made since 1990.

19.2 IO

19.2.1 Introduction and Early Studies

Our knowledge of Io's atmosphere has undergone a major revision in the last decade. Before 1990, observational information was restricted to several clear but indirect pieces of evidence, a single direct infrared detection by *Voyager* in 1979, and a number of upper limits from ultraviolet (UV) spectroscopy. Even loosely constrained, Io's atmosphere was quickly recognized as bearing unique features among planetary atmospheres, the most prominent being its apparent spatial and temporal variability. The lack of data did not hinder theoretical studies on the horizontal, vertical and chemical structure during the 1980s. Since 1990, the direct detection of it from Earth and Earth orbit in different wavelength ranges has provided a much firmer basis for our perception of Io's atmosphere. After a brief introduction, we concentrate in this section on recent developments. The earlier studies are detailed in previous reviews by Johnson and Matson (1989), Lellouch (1996), Spencer and Schneider (1996), and Trafton et al. (1998).

The first definite evidence for an atmosphere around Io was obtained in 1973 with the *Pioneer 10* detection of relatively dense ionospheric layers above Io's surface near the terminator (Kliore et al. 1974, 1975). Preliminary estimates of the neutral atmosphere required to explain these data yielded surface pressures of $10^{-8}$-10^{-9} bars. Shortly after, optical observations detected atomic sodium around Io (Brown 1974), and it was quickly established that the observed sodium formed a cloud of atoms in orbit around Jupiter that had escaped non-thermally from Io, implying a source of Na in Io's atmosphere or at the surface. Further evidence for atmospheric escape was obtained from the optical detection of a potassium cloud (Trafton 1975) and of ionized sulfur in the jovian magnetosphere (Kupo et al. 1976).

The “watershed event” for Io's atmosphere occurred the
same year with a triple discovery: the presence of active volcanism on Io’s surface (Morabito et al. 1979), the attribution of a 4.1 μm feature in Io’s infrared (IR) spectrum to solid SO₂ (Panale et al. 1979, Smythe et al. 1979), and the detection of gaseous SO₂ at 7.3 μm over the volcanic center Loki Patera (Pearl et al. 1979). The latter was interpreted as indicating a 100 nbar local atmosphere at 130 K, although a subsequent reinterpretation (Lellouch et al. 1992) has shown it to be consistent with lower pressures (5–40 nbar) and higher temperatures (up to 400 K). After this single observation, Io’s SO₂ atmosphere eluded detection for another eleven years, although several attempts in the UV were useful in placing upper limits on the global SO₂ abundance, the most significant being that of Ballester et al. (1990) which placed an upper limit of 2 × 10ⁱ⁷ cm⁻² on a homogeneous SO₂ atmosphere.

After the Voyager flybys and until the detection of SO₂ rotational line emission by Lellouch et al. (1990), theorists had only three data sets to characterize the atmosphere of Io: the two Pioneer 10 ionospheric radio occultation profiles and the detection of a volcanic plume over Loki by Voyager IRIS. A number of questions were raised: (1) how representative was the 100 nbar plume atmosphere of the whole atmosphere and how far would a plume atmosphere propagate horizontally? (2) was the atmosphere collisionally thin or collisionally thick to the penetration of thermal ions in the Io plasma torus as they swept by Io (if in the former, the atmosphere might be sputter generated)? and (3) did the Pioneer 10 detection of “dayside” and “nightside” ionospheres imply a global atmosphere but with substantial day/night surface pressure variations? Since most of the SO₂ gas from volcanic vents eventually condenses on the surface, what fraction of Io’s atmosphere was “buffered” by vapor pressure equilibrium with surface SO₂ frost/ice? These questions were pursued by modelers with limited success due to insufficient knowledge of the atmosphere, although two pieces of work in this period (McGrath and Johnson 1987, Schneider et al. 1991) did provide indirect evidence for an atmosphere with exobase above the surface (i.e., collisionally thick).

19.2.2 Recent Progress: Observations

SO₂ Atmosphere

Since the first detection of the SO₂ atmosphere via IR spectroscopy in 1979, two additional techniques have become important in characterizing the atmosphere in greater detail. The first is millimeter-wave spectroscopy, which detects SO₂ in emission. The second is UV observations (imaging and spectroscopy), which detect the SO₂ gas primarily in absorption. Advances in IR techniques have also enabled new SO₂ measurements from ground-based telescopes. We briefly describe the advances made since 1990 using these three techniques; results are also summarized in Table 19.1.

Millimeter Observations The first post-Voyager detection of SO₂ gas was made by Lellouch et al. (1990) using millimeter-wave heterodyne spectroscopy. The observations have been performed frequently since the initial detection, and yielded useful data in 1991, 1993, 1994, 1995, 1999, and 2002. To avoid contamination by Jupiter, the observations are conducted only near eastern or western elongation (orbital longitude L = 90° or L = 270° respectively), and they have low temporal (i.e., longitudinal) resolution and do not resolve the satellite spatially (10–20° spatial resolution). Twelve SO₂ emission lines, which span a factor of ~20 in line intensity, have been detected, all but one with relatively low energy levels (8 to 165 cm⁻¹). Because the rotational levels have very short collisional relaxation times, the observed transitions represent local thermodynamic equilibrium (LTE) thermal emission of the atmosphere at number densities as small as ~2 × 10⁶ cm⁻³ (Lellouch et al. 1992). The strongest have brightness temperature contrasts of 20–40 K above the surface continuum, implying that the mean dayside SO₂ gas temperature is higher than the mean surface brightness temperature by at least 20–40 K, perhaps much more if the dayside atmosphere covers only a fraction of Io’s surface and/or if the lines are not optically thick. The observed lines are fully resolved. The linewidth (FWHM) of the strongest lines is ~600 kHz at 220 GHz and scales as the line frequency, indicating Doppler broadening. Collisional broadening would imply an implausibly 10⁻⁴ bar surface pressure. The FWHM/frequency line ratio of ~2.7 × 10⁻⁸ gives a temperature of 910 K for thermal broadening, or a velocity of 0.8 km s⁻¹ for bulk velocity broadening. A sampling of the millimeter detections is shown in Figure 19.1.

The first interpretation (Lellouch et al. 1990, 1992) assumed that Io’s SO₂ atmosphere is in hydrostatic equilibrium, in which case T = 910 K is an upper limit to the mean atmospheric temperature. Since the bulk of Io’s atmosphere is likely to be at a much lower temperature (see discussion of radiative models below), the linewidths were interpreted as being affected by saturation effects. For reasonable values of the atmospheric temperatures, this indicated local column densities of 0.5–5 × 10¹⁷ cm⁻² covering only a small fraction of the disk. A more precise characterization requires multiline observations, whereby the relative contrast of at least two lines with known intrinsic strengths constrains their respective saturation degree, thereby helping to disentangle the temperature/pressure/coverage variables. This method, which motivated most of the SO₂ millimeter-wave observations over 1991–1999, is not equivalent to a rotational diogram analysis because all lines have similar lower energy levels. The line optical depth can be derived from the data, yielding the SO₂ column density (or equivalently the surface pressure). Once this is obtained, the extent of the projected surface (disk) covered by the atmosphere, fₚ, is inferred from the absolute line contrast. Converting this to a hemispherical coverage fₚ requires knowing how the gas is distributed. If the atmosphere is assumed to be restricted to a circular region around disk center, i.e., close to the subsolar point, fₚ and fₚ are related through fₚ = 1 – (1 – fₚ)²/².

In 1991, the non-detection of the weaker 146.605 GHz line on the trailing side led Lellouch et al. (1992) to conclude that the detected 222.965 and 143.057 GHz lines had a moderate optical depth (~2). This in turn implied a hot (Tₚ = 500–600 K), dense (3–15 nbar) and very localized (fₚ = 5–8%, fₚ = 2.5-4%) atmosphere. The high temperature and low areal extent for the trailing atmosphere were confirmed from the 1993 and 1994 observations (see Lellouch 1996). By contrast, the detection of the 146.605 GHz line on the leading side in March 1993 and May 1994 suggested a substantially lower temperature (250–400 K) and higher extent (fₚ = 12–
Table 19.1. Summary of SO₂ observations of Io since 1990.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Type</th>
<th>λ Coverage</th>
<th>Δλ</th>
<th>Date</th>
<th>Location</th>
<th>Spatial Fractional coverage Uniform coverage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lellouch et al.</td>
<td>1992</td>
<td>1.3-2 mm</td>
<td>100 kHz</td>
<td>1992</td>
<td>Trailing</td>
<td>none (1.5-7.5) x 10^{17} 500-600 2.5-4</td>
</tr>
<tr>
<td>Lellouch et al.</td>
<td>1996</td>
<td>1.3-2 mm</td>
<td>100 kHz</td>
<td>1993-94</td>
<td>Leading</td>
<td>none (1.5-2.5) x 10^{17} 250-400 6-9</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.8-2 x 10^{17} (250)^b 11-18</td>
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<td></td>
<td></td>
<td>(2.5-7.5) x 10^{17} 600 2.5-4</td>
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<td></td>
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<td></td>
<td></td>
<td></td>
<td>2 x 10^{16} (250)^b 29-36</td>
</tr>
<tr>
<td>Lellouch et al.</td>
<td>2000</td>
<td>1.3-2 mm</td>
<td>100 kHz</td>
<td>10/99</td>
<td>Lead.+Trail.</td>
<td>none 1.6 x 10^{17} 300 10</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>8 x 10^{16} (250)^b 20</td>
</tr>
<tr>
<td>Ballester et al.</td>
<td>1994</td>
<td>1900-2300 Å</td>
<td>4.5Å</td>
<td>03/92</td>
<td>none 1.6 x 10^{17} 300 10</td>
<td></td>
</tr>
<tr>
<td>Sartoretti et al.</td>
<td>1994</td>
<td>2325Å ±170Å</td>
<td>03/92</td>
<td>L = 285°</td>
<td>none 2 x 10^{19} (213) 10</td>
<td></td>
</tr>
<tr>
<td></td>
<td>1996</td>
<td>2600Å ±182Å</td>
<td>06/93</td>
<td></td>
<td></td>
<td>&gt; &lt; 4 x 10^{16} (213) 10</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2850Å ±240Å</td>
<td>07/93</td>
<td></td>
<td></td>
<td>(7) Clarke et al. 1994</td>
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<td></td>
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<td></td>
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<td></td>
<td></td>
<td>(9) Spencer et al. 1997</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>3.7 x 10^{17} (t) if 1 x 10^{19} (213) 10</td>
</tr>
<tr>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>τ is all SO₂ gas</td>
</tr>
<tr>
<td>Hendrix et al.</td>
<td>1999</td>
<td>2100-3200Å</td>
<td>13.7Å</td>
<td>05/98</td>
<td>L = 120°-150°</td>
<td>1 x 10^{19} (213) 25</td>
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<td></td>
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<td></td>
<td>4 x 10^{17} 4</td>
</tr>
<tr>
<td>McGrath et al.</td>
<td>2000a</td>
<td>1900-2300 Å</td>
<td>1.5Å</td>
<td>8/96</td>
<td>Pele Ra</td>
<td>3.25 x 10^{16} (213) 280</td>
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<tr>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>~750 km</td>
<td>1.5 x 10^{16} (213) 150</td>
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<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>7 x 10^{15} (213) 200</td>
</tr>
<tr>
<td>Spencer et al.</td>
<td>2000</td>
<td>2740-3365Å</td>
<td>3Å</td>
<td>10/99</td>
<td>Pele plume</td>
<td>7 x 10^{16} (t) (213) 21</td>
</tr>
<tr>
<td>Feldman et al.</td>
<td>2000a</td>
<td>1216Å Monochromatic</td>
<td>10/97</td>
<td>L = 258°</td>
<td>~200 km 3 x 10^{16} (213) 280</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.4 x 10^{16} (213) 150</td>
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<td></td>
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<td></td>
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<td></td>
<td>4.5 x 10^{16} (213) 200</td>
</tr>
<tr>
<td>Jessup et al.</td>
<td>2004</td>
<td>2050-3150Å</td>
<td>6.5Å</td>
<td>11/01</td>
<td>L = 152°, ±60° lat, ~200 km 3 x 10^{16} (213) 280</td>
<td></td>
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<td></td>
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<td>centered on</td>
<td>1.4 x 10^{16} (213) 150</td>
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<td></td>
<td>Prometheus</td>
<td>4.5 x 10^{16} (213) 200</td>
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<td>~200 km 3 x 10^{16} (213) 280</td>
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<td>~few x 10^{17}, (0-20)° lat 100-450</td>
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<td>~1 x 10^{16}, 100-450</td>
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<td></td>
<td></td>
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<td></td>
<td>(40-50)° lat</td>
<td></td>
</tr>
<tr>
<td>Spencer et al.</td>
<td>2002</td>
<td>18.9 µm λ/Δλ=50,000</td>
<td>11/01</td>
<td>L = 16°</td>
<td>none 2 x 10^{16} (213) 21</td>
<td></td>
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<td></td>
<td>L = 213°</td>
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<td></td>
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<td>1 x 10^{17} (213) 21</td>
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</tbody>
</table>

Note that ( ) under T indicates that the temperature is *assumed*, not derived or inferred. *Hydrostatic model; Plume interpretation; T = 250 K is assumed; I = imaging; S = spectroscopy; (t) = tangential.
16%, $f_0 = 6-9\%$) for Io's leading atmosphere. For all the observations recorded in 1993 and 1994, the hemispheric-average column is in the range $(1-2) \times 10^{16} \text{ cm}^{-2}$, with a tendency for higher columns on the trailing side than on the leading. Thanks to an unprecedented signal-to-noise ratio (S/N), the October–November 1999 observations resulted in the detection of many weak lines. Although not yet fully exploited, these observations confirm a cooler and more extended atmosphere on the leading side (Lellouch et al. 2000).

Best fits of the October 1999 data are obtained for $T_{\text{atm}} = 200$ K and $f_0 = 24\%$ on the leading, vs. $T_{\text{atm}} = 400$ K and $f_0 = 8\%$ on the trailing. The November 26, 1999 observation constitutes the best data set to constrain the gas temperature, as two close SO$_2$ lines were detected in the same spectrometer, eliminating possible calibration errors. This indicated $T_{\text{atm}} = 400 \pm 100$ K, again on the trailing side. However, the January 2002 observations included the high-energy (404 cm$^{-1}$) 265.481 GHz SO$_2$ line, which indicated a rotational temperature of only $180 \pm 60$ K (Lellouch et al. 2003).

The high temperatures inferred from the millimeter observations on the trailing side using hydrostatic models are at odds with radiative-conductive models (Strobel et al. 1994 – see Section 19.2.3) which predict that the atmosphere never warms above 200 K in the first scale height. This may suggest that the hydrostatic interpretation of the millimeter data is incorrect. Ballester et al. (1994) first proposed that the millimeter linewidths primarily reflect velocity dispersion within gaseous plumes rather than a combination of temperature and saturation effects. Lellouch (1996) and Lellouch et al. (1996, 2000) presented simplified models based on this idea. The hemispheric-average columns of $(0.6-2.5) \times 10^{16} \text{ cm}^{-2}$ achieved by these models are comparable to those in the hydrostatic models, but the data can now be fit even with low temperatures, and therefore the atmosphere is no longer necessarily "hot and localized". However, because the plumes are small (e.g., $r = 135$ km for an ejection velocity of $\sim 0.5$ km s$^{-1}$ as indicated by the data), they must be very numerous (50–300) to cover a significant fraction of one hemisphere. This large number may be somewhat decreased if allowance is made for a non-zero horizontal flow which increases the plume size. This number can be reduced further if an admixture of small and Pele-class plumes is assumed. With the $\sim 100$ active volcanic centers observed by Galileo (see Chapter 14 and map of Io in Appendix 1), 50 active plumes may not be unreasonable, especially if many of them are the invisible "stealth" plumes (i.e., those with a low condensate content) postulated by Johnson et al. (1995).

The possible existence of almost purely gaseous plumes has been demonstrated by Kieffer (1982) in the case of a high-entropy erupting fluid from a reservoir of superheated SO$_2$ vapor in contact with a deep, hot and dense silicate melt (1400 K, 40 bar). While already complex to implement, the plume models are certainly far from Io's reality, as stressed by Lellouch (1996).

**Ultraviolet Observations** SO$_2$ gas absorbs strongly in the UV wavelength region, as shown by the absorption cross section in Figure 19.2. This figure and Table 19.1 summarize the numerous successful UV observations made since the first UV images of Io (Paresce et al. 1992) and the first successful UV spectroscopic detection of SO$_2$ in absorption (Ballester et al. 1994). The initial spectroscopic observations (Ballester et al. 1994, Clarke et al. 1994, Trafton et al. 1996) had no spatial resolution and only modest spectral resolution (see Table 19.1). Unlike the millimeter-wave spectrum, the UV observations are primarily sensitive to the column-integrated abundance of the absorbing gas, but
not to its temperature, except for a general decrease in the band contrast with increasing temperature and subtle variations in the band peak position and skewness (Wu et al. 2000). Analysis is subject to two complications, which make unique interpretation of the data difficult. The first comes from the fact that the SO$_2$ UV spectrum has a very complex structure of many densely packed lines that have not been resolved in laboratory measurements, so that line-by-line position and intensity information is not available. In this situation, applying Beer’s law at a spectral resolution comparable to that of the measurements can lead to significant underestimates of absorber abundance (Belton 1982).

Band models are much preferred, and several modelers have followed the treatment by Ballester et al. (1994) based on a Malkmus intensity distribution of lines with Lorentz profiles and including variations with temperature. The other complication is due to the poorly known contribution of Io’s surface to the overall geometric albedo. SO$_2$ frost, a known constituent of Io’s surface, has broadly similar spectral properties as the gas absorption, making the competing optical effects of gas and frost very difficult to disentangle. The reflectance depends sensitively on the frost grain size and how it is physically mixed with the other components. While it is known that SO$_2$ frost is dark in the UV, it is impossible to reliably predict the absolute surface reflectance and its spectral dependence. As a consequence, only observations with spectral resolution high enough to distinguish characteristic gas spectral features unambiguously constrain gas abundances.

For example, while it is possible to model the early UV imaging data of Sartoretti et al. (1994, 1996) purely in terms of variations of surface properties, an alternate explanation is that the darkest UV component seen in the images represents patches of SO$_2$ gas. In their preferred model, Sartoretti et al. (1996) assume that the frost reflectivity is at the lowest value allowed by their data ($R = 0.013$ at 2000 Å), which implies the presence of SO$_2$ patches with typical columns of $\sim 1 \times 10^{15}$ cm$^{-2}$ covering 11–15% of the projected surface. One of these patches is Pele, and most others are well correlated with known thermal anomalies (Veeder et al. 1994). However, it is important to emphasize that these early UV images were incapable of detecting SO$_2$ column densities $\lesssim 8 \times 10^{16}$ cm$^{-2}$, so the presence of a more uniformly distributed, lower-density component could not be ascertained.

The lack of spatial resolution in the disk-average spectroscopic observations (e.g., Ballester et al. 1994; Clarke et al. 1994; Trafton et al. 1996), when coupled with the ambiguity in surface reflectance, permits two different model domains to fit the data equally well. The trade-off between spatial coverage and atmospheric column abundance leads to equally good fits with either spatially confined regions with relatively high SO$_2$ columns, or much more extended coverage with more modest SO$_2$ column densities. The best fit solutions for both domains are summarized in Table 19.1. The hydrostatic interpretation of the millimeter observations and some of the early UV observations (Trafton et al. 1996, Sartoretti et al. 1994, 1996) showed a trend to more SO$_2$ gas on the trailing as opposed to the leading hemisphere.

Another important constraint for the UV spectroscopy is the lack of detection of SO$_2$ bands in the near-UV ($\lambda \geq 2500$ Å). Clarke et al. concluded that this absence of fine structure ruled out a global atmosphere denser than $4 \times 10^{16}$ cm$^{-2}$. However, they pointed out that a very dense, localized component (e.g., $2 \times 10^{19}$ cm$^{-2}$ over 10% area), was not inconsistent with the data. At these very high columns, while structure appears in the region of minimum absorption near 2350 Å, the 2800–3100 Å region is saturated out to 100% absorption, consequently showing no spectral contrast. In addition Hendrix et al. (1999) found a clear decrease of the albedo shortward of 2360 Å (see Figure 19.3) which cannot be explained by SO$_2$ frost. Their interpretation of this decrease as due to SO$_2$ gas absorption, and not surface reflectance effects, led them to infer very large $(10^{19}$ cm$^{-2}$) SO$_2$ column densities over at least 25% of the surface.

As both spatial and spectral resolution have improved, spatial variability of the SO$_2$ atmosphere has begun to be detected directly and quantified. McGrath et al. (2000a) obtained spatially resolved spectroscopy at 1.5 Å spectral resolution using a small aperture to target three specific loca-
tions on Io’s disk, one of which was Pele. [The importance of spectral resolution adequate to distinguish between gas and frost is illustrated in Figure 19.3, where their data are compared with that of Ballester et al. (1994) and Hendrix et al. (1999).] The targets were chosen to sample different physical conditions that are likely to exist on Io’s surface: (1) the Pele volcano; (2) Ra, a region bright in the visible and dark in the UV, indicating abundant SO$_2$ frost; and (3) a reference region at 45°S and 300°W (“T3”) that is dark in visible and bright in UV, i.e., presumably frost-poor. They derived an SO$_2$ column density of only 3.25 x 10$^{16}$ cm$^{-2}$ at Pele (results for the other targets are given in Table 19.1).

The observation of T3 was a strong indication of a spatially variable but widespread atmosphere; it is particularly significant to detect SO$_2$ in a region where no active plume has ever been observed, although this region (Aten Patera) is close to the site of a known hot spot (Lopes-Gautier et al. 1999) and seems to have experienced a Pele-like eruption.

Spencer et al. (1997) imaged the Pele plume against dark sky and silhouetted against Jupiter’s disk during Io’s transit only 7 days later, showing Pele to be active. Pele’s plume height was measured to be 420 ± 40 km, and its opacity at 2720 Å to be ~0.19. The plume was not detected at 3400 and 4100 Å. This wavelength-dependent optical depth was interpreted as due to absorption by either small dust particles (<0.08 μm) with a total mass of at least 1.2 x 10$^9$ g, or SO$_2$ gas with a column abundance of ~3.7 x 10$^{17}$ cm$^{-2}$. Note that the required mass for SO$_2$ gas is as much as 100 times larger than for dust, so this interpretation would clearly classify Pele as a “stealth plume.”

The order of magnitude difference in SO$_2$ column between the Spencer et al. (1997) and McGrath et al. (2000a) results was at first thought to be due to particulates accounting for some of the opacity in the UV images. Even so, these SO$_2$ columns were significantly less than the 10$^{18}$–10$^{19}$ cm$^{-2}$ discussed by Sartoretti et al. (1994, 1996) and Hendrix et al. (1999). When Spencer et al. (2000) obtained both imaging and spectroscopy of the Pele plume again in 1999 they made the spectacular discovery of gaseous S$_2$ through 15–20 bands belonging to the B$^2$Σ$_u^+$–X$^2Π$$_g^-$ system (Figure 19.4), in addition to a detection of SO$_2$ gas at shorter wavelengths. The S$_2$ provides a significant source of opacity in the 2500–3000 Å region previously attributed to SO$_2$. When their tangential SO$_2$ column density of ~7 x 10$^{16}$ cm$^{-2}$ is converted to a vertical column (~factor of 2 decrease) it is in remarkably good agreement with the ~3.25 x 10$^{16}$ cm$^{-2}$ column found by McGrath et al. (2000a). In retrospect, the detection of S$_2$ and SO$_2$ in the Pele plume at much lower abundance than inferred by Sartoretti et al. (1996) and Spencer et al. (1997) indicates that the most significant source of opacity in their images may have been due to absorption by gaseous S$_2$, with negligible dust extinction and only a minor contribution due to gaseous SO$_2$.

Further evidence in support of the lower density, larger coverage regime has come from HI Lyman-α images of Io obtained by Roesler et al. (1999), which has opened a new avenue for studying the distribution of Io’s SO$_2$ atmosphere. Such observations have been acquired in 1997, 1998, 1999, 2000, and 2001, although only the 1997 and 1998 observations have been published to date (Figure 19.5). In the “1998 West” observation (see Figure 19.5) the Lyman-α emission appears to consist of two mid-to-high latitude (~25°) bright patches at the 2 kR level, while the equatorial latitudes are dark, with ~0.7 kR at disk center, and longitudinal and/or temporal variations. The brightest Lyman-α emission is entirely within the disk. Although Roesler et al. proposed several explanations for the observed structure, differential absorption of surface reflected solar Lyman-α photons by low latitude atmospheric SO$_2$ was shown to be the most likely by Feldman et al. (2000b) and Strobel and Wolven (2001). Since the SO$_2$ cross section is large at Lyman-α ($σ = 3.9 x 10^{-17}$ cm$^2$) SO$_2$ must be a significant absorber for typical SO$_2$ columns of several 10$^{16}$ cm$^{-2}$. This hypothesis was confirmed by 1999 Lyman-α images taken during and following eclipse. No polar bright regions are apparent in the eclipse images, while they are present in contiguous post-eclipse images, which conclusively links the bright regions near the poles with sunlight.

The Lyman-α images thus provide a “negative image” of the SO$_2$ atmosphere, substantially clarifying its spatial structure at scales ~200 km. Although the problem of disentangling the atmospheric signature from the surface reflecting signature remains, the retrieval of absolute SO$_2$ columns from these images is facilitated by the fact that SO$_2$ is a continuum absorber at Lyman-α (enabling the use of Beer’s law). Assuming a reflectivity of 0.05 at the poles (which corresponds to a geometric albedo of 0.033), and that the reflectivity is uniform across the disk, Feldman et al. (2000b) use Beer’s law to determine the optical depth at the sub-Earth point, which corresponds to SO$_2$ column densities in the range (1–4) x 10$^{16}$ cm$^{-2}$. Strobel and Wolven (2001) constructed a spatial model of the Lyman-α emission for the “1998 West” image shown in Figure 19.5 based on longitudinally homogeneous model atmospheres with column densities decreasing sharply from (1–1.5) x 10$^{16}$ cm$^{-2}$ at the equator to ~3 x 10$^{14}$ cm$^{-2}$ at 50° and beyond. These models capture the essential observational features and suggest that Io’s atmosphere is restricted to an approximately ±30–40° band in which lateral inhomogeneities (at the resolution of the data) are modest. Strobel and Wolven (2001) interpreted this in the context of numerous (10–200) plume atmospheres, with a total vent ejection rate of ~5 x 10$^{30}$ molecules s$^{-1}$.
Finally, the most recent long-slit UV spectroscopy from HST has achieved the best spatial resolution to date (Jessup et al. 2004). The observations cover latitudes from 60° N to 60° S, and include the volcanoes Zamama, Malik, Tu­pan, and Chaac, as well as SO₂ frost plains. There is only a modest difference in the SO₂ gas column abundance be­tween the Prometheus plume and the nearby low-latitude SO₂ frost plains. Initial analysis shows a smooth transi­tion in SO₂ abundance with latitude, and a fall-off of the abundance at high latitudes, qualitatively consistent with both the Lyman-α imaging and McGrath et al. (2000a) results discussed above. Additionally, continuum emission near 2800 Å is observed over the Prometheus plume and neighboring low latitude regions. The continuum emission also appears to correlate with the SO₂ absorption, which peaks at Prometheus. Thus the concept of the SO₂ atmos­phere discussed above appears to be more complex than one dominated by scores of active volcanoes. These observations show only modest spatial variability of the SO₂ atmosphere, with relatively small enhancements over active volcanoes, consistent with the McGrath et al. (2000a) results.

**IR Spectroscopy** Ground-based disk-averaged IR ob­servations of Io’s atmosphere in November 2001 at a resolution of 50 000 (Spencer et al. 2002) resulted in the detection of 15 lines of the ν₃ band of gaseous SO₂ in the 18.9 μm region. The lines appear in absorption, probably due to non-LTE effects that produce absorption even for atmospheric kinetic temperatures somewhat warmer than the surface temperature. Line strength varied strongly with orbital longitude, being greatest on the anti-Jupiter hemisphere. A prelim­iary analysis of the data gives disk-averaged SO₂ column densities of ~2 x 10¹⁶ and 1 x 10¹⁷ cm⁻² at L = 16° and 213° respectively. They also attempted to detect the SO₂ ν₃ band at 7.3 μm seen by Pearl et al. (1979) over Loki. A 15­minute integration showed no absorption at sensitivity levels that should have detected the absorption strength seen by Pearl et al. (1979).

**Minor Compounds**

Beyond SO₂, a number of other compounds have been searched for in Io’s atmosphere. SO was first clearly detected in 1995 millimeter observations at 219.949 and 138.176 GHz on both sides of Io (Lellouch et al. 1996), and confirmed in 1999 at 251.825 and 109.252 GHz. The strongest SO lines appear typically two times weaker than the strongest SO₂ lines, although they are intrinsically five to ten times more intense, implying that SO is a minor compound. In the framework of hydrostatic models the observations cannot distinguish between a hemispheric SO atmosphere – in this situation, a barely collisionally thick SO atmosphere with a (2-6) x 10¹⁴ cm⁻² column is indicated – and an SO component colocated with SO₂ on a restricted fraction of Io’s surface with a 4-10% SO/SO₂ mixing ratio. In the case of volcanic models, the SO/SO₂ mixing ratio within the erupt­ing plumes is also in the range 3-10%. The McGrath et al. (2000a) data also show several weak features consistent with SO bands that give a relative mixing ratio to SO₂ of about 10%, consistent with the millimeter-wave detections.

Another important measurement of SO was made using infrared spectroscopy of Io during eclipse in 1999, which led to the detection of the forbidden electronic a¹Δ → X³Σ⁻ transition of SO at 1.71 μm (de Pater et al. 2002). The detection was later confirmed by Goguen and Blaney (2001). These emissions are attributed to gas from the volcanic vents Loki and Janus/Kaneheki for the two observations, respective­ly. The de Pater et al. (2002) observations indicate an emission rate of ~2 x 10²⁷ photons s⁻¹. They discussed many possible mechanisms for this emission and concluded it to be caused by direct ejection of SO molecules in the excited a¹Δ state from the vent at a quenching temperature of ~ 1500 K. The shape of the band indicates a rotational temperature of ~ 1000 K. Because rotational levels are easily thermalized, this temperature may represent the actual kinetic temperature of the emitting gas as it is vented.

As mentioned above, Spencer et al. (2000) discovered S₂ in the Pele plume (see Figure 19.4). Best fit models indicate an S₂ (tangential) column density of (1 ± 0.2) x 10¹⁶ cm⁻² and a temperature of 300 K. Since SO₂ is detected in the same observation, an SO₂/S₂ ratio of ~3-12 can be inferred. The discovery of S₂ was not unexpected, since sulfur vapor had been proposed to be the driver of the Pele plume (McEwen et al. 1988), but the apparent dominance of SO₂ gas was not predicted. Given the observed abundance, the equivalent sublimating rates by SO₂ and S₂ are 0.12 and 0.8 mm yr⁻¹ in the Pele plume deposits. The S₂ detection may be the key in explaining the red deposits near Pele and other active regions, as S₂ is unstable against photolysis, producing reddish S₃ and S₄ molecules by poly­merization. However, S₂ also reacts rapidly with O atoms (k = 2.2 x 10⁻¹¹ cm³ s⁻¹) and S reacts rapidly with O₂ (k = 2.2 x 10⁻¹² cm³ s⁻¹). To polymerize to S₃ and S₄ would require an environment with a low total O to total S ratio such as in the vicinity of Pele.

The first detection of gaseous NaCl was achieved in Jan­uary 2002 (Lellouch et al. 2003) via the detection of emission lines at 234.252 and 143.237 GHz. The disk-averaged column density is in the range (0.8-20) x 10¹³ cm⁻², with a preferred value of 4 x 10¹³ cm⁻², i.e., about 0.4% of SO₂. Because of its vanishingly low vapor pressure at Io’s temperature, the most likely source of NaCl is direct volcanic output, although sputtering of salt-bearing atmospheric aerosols is not excluded. Volcanic plume models indicate total volcanic emission rates of (2-8) x 10⁸ NaCl molecules s⁻¹, i.e., typically 0.3-1.3% of the SO₂ rates. NaCl is probably restricted to smaller regions than SO₂ because of increased photolytic and condensation losses.

Eight other compounds (CO, H₂S, OCS, S₂O, ClO, CS,
NaOH, KCl) have been searched for unsuccessfully at millimeter wavelengths. The most significant of the associated upper limits is probably the stringent $10^{-16}$ bar upper limit on a global H$_2$S atmosphere (Leboulleux et al. 1992). An upper limit of $2 \times 10^{14}$ cm$^{-2}$ was set for the abundance of CS$_2$ by McGrath et al. (2000a).

Atomic Species

Five atomic species have been detected in emission to date in Io's atmosphere: O, Na, K, and Cl. O, S, and Cl are detected via collisional excitation of neutrals in the atmosphere and corona by plasma torus electrons, while Na and K are observed primarily via their strong resonant fluorescence transitions at visible wavelengths. Atomic sulfur and oxygen have been observed extensively both far from Io in the plasma torus since 1981 (Brown 1981, Durrance et al. 1983), and near Io since 1986 (Ballester et al. 1987). A current unresolved issue is whether the predominant mechanisms for the collisionally excited emissions near Io is direct excitation of atomic species, or dissociative excitation of molecular species. Observations of these emissions are not straightforward to interpret as regards atmospheric properties because they represent an integrated brightness along the line of sight that is a product of the electron and neutral densities and the appropriate rate coefficient, itself a function of electron temperature. The integrated brightness is therefore diagnostic of both the plasma conditions and the relevant neutral densities, and information about the two are difficult to disentangle unambiguously. A common approach used to estimate neutral column densities is to assume that the excitation process is electron impact of atomic species, and that both the electron density and temperature are constant along the line of sight, both of which are poor approximations. The plasma properties near Io are complex (Linker et al. 1998, Combi et al. 1998, Saur et al. 1999), and currently forward modeling, as opposed to inversion, is more successful at deriving information about the atmosphere (cf. Saur et al. 2000). Interpretation is also complicated by variations due to observing geometry and the location of Io relative to the plasma torus equator. Finally, especially for atomic sulfur, there is very limited information available about the relevant rate coefficients. Hence although there are extensive observations of the atomic emissions, such as those of Oliversen et al. (2001), relatively little information about atmospheric properties has been extracted from them. We provide here a brief summary of the atmospheric diagnostics; the derived plasma properties are described in more detail in Chapter 22.

Electron temperature and column density estimates for O and S were first made by Ballester (1989) using disk-average International Ultraviolet Explorer (IUE) spectra. The oxygen emissions were found to be consistent with the excitation of oxygen in Io's exosphere by nominal ~5 eV torus electrons. Minimum oxygen column densities of $N_O > (4-7) \times 10^{13}$ cm$^{-2}$ were required to produce the emissions assuming canonical torus values of $T_e = 5$ eV and $n_e = 2000$ cm$^{-3}$. Employing information about the optical depths of the allowed and forbidden transitions of atomic sulfur, Ballester (1989) also derived limits on the sulfur column density of $2.2 \times 10^{12}$ cm$^{-2} < N_S < 7 \times 10^{15}$ cm$^{-2}$. In the spatially resolved spectroscopic observations of McGrath et al. (2000a) described earlier, emission from the SI 1900,1914 Å doublet was detected in two of their three targets, with intensities of 3.6 kR over Pele and 1.6 kR over T3. The sulfur column above Pele was estimated to be $N_S \sim 1 \times 10^{14}$ cm$^{-2}$ using techniques similar to Ballester (1989). Using HST/STIS data, Feaga et al. (2002) were able to resolve the optically thick allowed and optically thin forbidden sulfur transitions at 1479 Å, which provided a significantly improved estimate of the tangential sulfur column density of $3.6 \times 10^{12}$ cm$^{-2} < N_S < 1.7 \times 10^{13}$ cm$^{-2}$ that is independent of electron density and temperature. [Using more recent atomic data for sulfur has now raised this upper limit to 1.3 $\times 10^{14}$ cm$^{-2}$.] Using the 1-D photochemical model of Summers and Strobel (1996), and assuming spherical symmetry, the vertical column density would be about a factor of 7 lower.

The observations that produced the Lyman-α images discussed above have also provided simultaneous monochromatic images of the far-UV S and O emissions. Along with Galileo broadband SSI images (Geissler et al. 1999) these data have revealed the complex morphology of the emissions, which is characterized by five notable features: equatorial "spots", volcanic plume glows, a limb-brightened ring of emission just off the disk, diffuse atmospheric emissions (also referred to as "glow"), and emission from Io's extended corona. The spots (often referred to as the "Io aurora") are observed to rock about the equator in concert with the changing orientation of the background jovian magnetic field, which provides a powerful tool to infer properties of the strong electrodynamic interaction between plasma and satellite. The limb-brightened rings of sulfur and oxygen emission imply that both species form global components of the atmosphere.

The Galileo SSI images of Io (Geissler et al. 1999) were acquired over the course of 14 eclipses from 1996–1998. While most of the observations were made using the SSI clear filter (wavelength coverage of 3800–10 400 Å), several sequences included imaging with the violet (3800–4450 Å), green (5100–6050 Å) and red (6150–7100 Å) filters. The equatorial spots were seen with all of the filters, but were brightest in the violet filter images. The diffuse atmospheric glow was seen against the disk of Io in the green filter, and particularly on the nightside in the E15 orbit observations. The identity of the emitters could not be unambiguously determined because of the broad wavelength coverage of the filters, however, several candidates were proposed by Geissler et al. (1999), including [OI] 6300 and 6360 Å, Ho 6560 Å, and SII 6720,6730 Å in the red filter; [OI] 5580 Å and NaI 5890,5900 Å in the green filter; and molecular emission from SO$_2$ in the violet filter. High resolution Keck spectra obtained by Bouchez et al. (2000) detected auroral emission from [OI] 6300,6363,5577 Å, and NaI 5889,5896 Å. They concluded that the red filter emissions seen by Galileo SSI were from the oxygen lines, and the green emissions were from the Na lines. They detected no emissions in the SSI violet region, lending support to the idea that the Galileo violet emissions are from broadband continuum emission from SO$_2$ or SO.
spatial resolution of $\sim 0.05 R_{\oplus}$ out to distances of $\sim 10 R_{\oplus}$. The coronal emission profiles vary considerably in slope and intensity, and are generally brighter when Io is on the dusk side of Jupiter. The intensities of emission from regions both near Io and in the extended corona vary with System III longitude in a near-simultaneous fashion, suggesting torus electron density as the probable source of this modulation. The observed ratio of oxygen to sulfur emission is relatively constant in time, perhaps reflecting the stoichiometric ratio of the SO$_2$ source molecules. A dramatic increase in profile brightness and slope between eclipse and post-eclipse observations in February 2000 suggests a dynamic response by a sublimation-supported component of Io's SO$_2$ atmosphere and associated atomic species.

It has long been believed that Io's SO$_2$ atmosphere condenses on to the surface when Io goes into eclipse. Several observations bear on this question. Visible emissions were seen during eclipse by Voyager 1 and were thought to be from SO$_2$ (Cook et al. 1981). Disk-averaged observations of Io passing into Jupiter shadow acquired by Clarke et al. (1994) showed that the far-UV sulfur and oxygen emissions, which had typical brightnesses of 1 kR when Io was in sunlight, decreased to only a few hundred Rayleighs within 20 min of Io entering eclipse. Geissler et al. (1999) report on the changes seen in a set of two SSI images obtained 11 min after the start of eclipse and 41 min later in orbit E15. Io's disk clearly darkened as the eclipse progressed, while a difference image shows that the plume glows have brightened (see their Fig. 2). Wollen et al. (2001) report on HST/STIS observations obtained on February 25, 2000 which show a dramatic increase in S and O brightnesses after eclipse egress. Retherford (2002) quantified these changes for the spots, the limb glow, and the extended corona. A thorough discussion of the timescales of various changes in the atmosphere (both atomic and molecular) associated with eclipse ingress and egress is presented by Retherford (2002; see his Chapter 5, especially Table 5.2). He concludes that after ingress, the sublimation atmosphere will collapse on a timescale of $\sim 5$ min, followed by a depletion of the atomic atmosphere within (conservatively) $\lesssim 30$ min. The atomic corona, however, is depleted on a timescale of $\sim 280$ min, which is longer than the duration of an eclipse ($\sim 130$ min), consistent with the STIS eclipse observations.

Cl has been detected with HST (Retherford et al. 2000, Feaga et al. 2004), however, the lack of adequate S/N and reliable electron excitation rate coefficients have hindered interpretation of the data, and only poorly constrained estimates of the abundance are available. Preliminary analysis indicates that Cl may be present in amounts comparable to that of Na (Feaga et al. 2004).

Because they have very large oscillator strengths and are therefore easily detected even at very low densities, observations of visible wavelength resonance fluorescence lines of sodium and potassium have been very important in studies of Io's atmosphere. In fact, much of what we know about the atmospheric dynamics and Io's interaction with the plasma torus comes from sodium observations. Mutual eclipses between Io and other Galilean satellites (Schneider et al. 1991, Burger et al. 2001) have allowed observations of Io's corona inside the Hill sphere ($\sim 8 R_{\oplus}$), down to $\sim 1.4 R_{\oplus}$, from which radial profiles of Na column density have been derived. Burger et al. (2001) have also shown that the Na corona is denser on the sub-Jupiter side, with the average radial profile $N_{Na}(b) = 2.2 \times 10^{12} b^{-2.34}$ for $b > 1.5 R_{\oplus}$. Bouchez et al. (2000) eclipse spectroscopy detected Na emission, most likely excited by torus electrons, from which they derived a disk-average column density (assuming nominal torus conditions) of $N \sim 4 \times 10^{12} \text{cm}^{-2}$, which is comparable to the Na column density derived by extrapolating the Schneider et al. (1991) and Burger et al. (2001) profiles to the surface. Potassium measurements have only been made down to $\sim 10 R_{\oplus}$ due to the weaker emission intensity, giving column densities at that distance of $\sim (0.4-1.5) \times 10^{10} \text{cm}^{-2}$. Contemporaneous measurements of Na allow the Na/K ratio to be derived, and it is approximately constant from $\sim 10-20 R_{\oplus}$, at a value of $10 \pm 5$ (Brown 2001).

The source of alkalis to Io's atmosphere is still not certain, in part because the observations are principally of escaped sodium in extended clouds rather than in the bound atmosphere near the surface. The salty satellite models (Fenale et al. 1974, Kargel 1991, Zolotov and Shock 2001) suggest that sodium might occur as a sulfate. The presence of Cl in amounts comparable to Na may imply that the initial volcanic source contains NaCl, although Cl could occur in other molecular forms (Zolotov and Shock 2001). The recent detection of NaCl indicates that this gas is a quantitatively plausible source of sodium and alkalis in Io's atmosphere. Since NaCl from a volcanic source efficiently dissociates, the resulting Na and Cl would react separately with the principal surface constituents, frozen SO$_2$ and its radiation products. Chlorine was therefore suggested to be present on the surface as Cl$_2$SO$_2$ (Schmitt et al. 2001) and Na as Na$_2$SO$_4$ (Wiens et al. 1997) or Na$_2$S$_2$ (Chrisey et al. 1988).

The vapor pressure of likely sodium-bearing molecules is low at the ambient surface temperatures on Io, leading Matson et al. (1974) to suggest initially that the source of the observed sodium was sputtering by the plasma trapped in Jupiter's magnetosphere. The corotating plasma ions do not reach Io's surface due to its ionosphere and collisionally thick SO$_2$ atmosphere. However, estimates of the energetic ion flux that does reach the surface suggest it could provide an adequate source of surface sputtered sodium (Wong and Johnson 1996b). An alternate source is direct population of the atmosphere from vents and volcanic plumes. Laboratory studies show that energetic ions, as well as photons and electrons (Wiens et al. 1997, Chrisey et al. 1988, Madey et al. 1998), preferentially remove alkalis from refractory materials by an electronic sputtering process that also causes decomposition of such materials. The sodium ejected in this fashion is predominantly atomic, with a smaller fraction in molecular form. The discovery that a component of the torus is sodium-containing molecular ions (Wilson and Schneider 1994) led to estimates of the abundance of molecular Na at the exobase. Unlike SO$_2$, Na is mostly in atomic form near the exobase (see, e.g., Wong and Johnson 1995). The molecular ions are then supplied to the torus by ion-molecule reactions near the exobase (Johnson 1994), or by direct flow along field lines that connect with the ionosphere (Wilson and Schneider 1999, Wilson et al. 2003).

Sodium emission has also been seen far up and down stream from Io. In addition, a giant nebula of escaping sodium (referred to as the jovian xeno-corona) has been seen
out to ~500 jovian radii (Mendillo et al. 1990). These Na observations clearly indicate that Io’s atmosphere is escaping and the ejected material supplies the Io torus. They have been used to derive a detailed description of atmospheric loss (Smyth and Combi 1997, Wilson et al. 2003). This is discussed in more detail in Chapter 23. Smyth and co-workers (Smyth and Marconi 2000) have applied the ideas learned from escape of sodium to loss of the principal species from Io’s atmosphere.

Ionosphere

Only two sets of detections of Io’s ionosphere have been made in the past 30 years, the first by the Pioneer 10 spacecraft, and the second by the Galileo spacecraft. A useful summary of the Pioneer 10 observations and their interpretation, which met with only limited success, is given by Johnson and Matson (1989). It is important to emphasize that the nature of the radio occultation measurements results in a viewing geometry that always puts both the entrance and exit measurements within a few degrees of the terminator. As a result, both the entrance and exit measurements primarily sample the sunlit atmosphere – even when they occur above the nighttime terminator, only the lower few km of the atmosphere is in darkness.

Results from a series of six Galileo radio occultation measurements in 1997, published in a single comprehensive paper by Hinson et al. (1998), have greatly clarified the situation with Io’s ionosphere. These occultations sampled a wide variety of geometries of the sunlit hemisphere relative to the plasma ram direction. The measurements yield information about the distribution and motion of the plasma near Io. The distribution was found to have two components. The first is present within a few hundred km of Io’s surface throughout the upstream and downstream hemispheres and resembles a bound ionosphere. Vertical electron density profiles for this component were derived at 10 locations near Io’s terminator. The peak density exceeded $5 \times 10^7 \text{ cm}^{-3}$ at 9 out of 10 locations, and reached a maximum of $2.8 \times 10^8 \text{ cm}^{-3}$. The peak density varied systematically with Io longitude, with maxima near the centers of the sub- and anti-jovian hemispheres, and minima near the centers of the downstream and upstream hemispheres. This pattern may be related to the Alfvénic current system induced by Io’s motion through the magnetospheric plasma (see Chapter 22). The vertical extent of the bound ionosphere increases from ~200 km near the center of the upstream hemisphere to ~400 km near the boundary between leading and trailing hemispheres.

The second component is highly asymmetric, consisting of a wake or tail that appears only on the downstream side and extends to distances as large as 10 Io radii. Plasma near Io’s equatorial plane is moving away from Io in the downstream direction, with velocity increasing from 30 to 57 km s$^{-1}$ between 3 and 7 Io radii. The latter velocity corresponds to corotation, suggesting that bulk plasma motion, rather than wave motion, was being observed. It is apparent that the major factor determining the morphology of the ionosphere is the plasma ram direction. The Galileo measurements generally confirm the original Pioneer 10 results, providing strong evidence, given the 23 year time span between measurements, that the ionosphere is stable. Most modeling of Io’s ionosphere to date has attempted to match the Pioneer 10 entrance profile, which is now clearly understood from Galileo measurements to be dominated by electrons on the flanks. The inability of the one-dimensional photochemical models to match this profile is therefore understandable.

Variability

Several sets of data show that Io’s atmosphere exhibits a relatively large degree of stability in a global sense: those of the ionosphere from Pioneer 10 and Galileo; the long-term Na observations (Schneider et al. 1991, Burger et al. 2001); and the permanent detectability of the SO$_2$ atmosphere. Nonetheless, several independent lines of evidence from the observations discussed above also suggest significant temporal variability of the atmosphere. The non-detection of IR absorption from the Loki region by Spencer et al. (2002) described above may indicate significant variability compared to the Voyager epoch, although a different interpretation is not precluded because of the lack of spatial resolution in the more recent observations. In some of the Lyman-α images the north polar bright region is almost absent, and several images covering approximately the same longitudes but at different times (cf. the 1997 image in Figure 19.5 is also “West” and can be qualitatively compared with the 1998 West image – see also Table 19.1 in Feldman et al. 2000a) show changes in the brightness and extent of the polar spots, indicating a greater or lesser latitudinal extent of the SO$_2$ atmosphere. The Strobel and Wolven (2001) analysis of the August 1998 Lyman-α image derived a low latitude column of ~$(1-2) \times 10^{16} \text{ cm}^{-2}$, in good agreement with the McGrath et al. (2000a) spectroscopic result for Pele and Ra at similar latitude, but a column of ~$3 \times 10^{14} \text{ cm}^{-2}$ for latitude $> 45°$ compared to the McGrath et al. (2000a) value of $7 \times 10^{15} \text{ cm}^{-2}$ at a latitude of 45° S, which is more than an order of magnitude larger. This would be consistent with the observed differences in the Lyman-α images mentioned above. Significant changes over relatively short time periods have also been seen in the Lyman-α images (e.g., October 1999), however, it is unclear if these changes are temporal or spatial, as the central longitude of Io also changed significantly during the course of the observation.

While the SO$_2$ millimeter emissions appear to be permanently detectable, this ensemble of data also provides evidence for temporal variability. In June 1995, the trailing side 143.057 GHz line appeared sharper and narrower than in previous years (see Fig. 8 of Lellouch 1996). In the framework of the hydrostatic equilibrium models, this suggests a reduction of either the column density or the temperature, and an increase in the areal extent. In fact, this observation, unlike all others, is consistent with a hemispheric atmosphere. In October 1999, the 221.965 GHz line was about 50% stronger than in 1990–1994, interpreted as a generally higher surface coverage. Beyond the leading/trailing differences discussed above, the 1999 observations allowed, for the first time, to search for orbital variations of strong line characteristics. While no obvious variability of the linewidth was found, there is a suggestion of an increase of the line area over $L = 40 – 135°$ and a decrease over $L = 240 – 340°$ (Figure 19.6a), consistent with the Spencer et al. (2002) 19 μm data. More definite is the variation of line frequency with
or thermal conduction, rotation, latent heat.

The early equilibrium models of Fanale, in their published papers, indicated that Io's atmosphere must contain significant amounts of SO, SO2, O, and S. These models produce slightly asymmetric and redshifted (100–200 m s−1) lines. This is consistent with the trailing side observations, but opposite to the leading side results. The interpretation might be related to angular momentum transfer from the plasma flow hitting Io's trailing side at ~57 km s−1. Saur et al. (2002) present evidence for torus ion drag forces compressing the ram/upstream atmosphere and extending the downstream atmosphere with a respective scale height ratio of 1/2. The ion drag forces will set the neutral gas in motion to yield a global blueshift on the leading/downstream side and a global redshift on the trailing/upstream side. We note that this type of gas motion has been deduced from the Galileo radio occultation measurements, which were described above.

Two additional pieces of evidence support the idea of temporal variability. A second set of Jupiter transit observations of Pele in 2003 do not show a strong S2 signature, as contrasted with the 1997 detection described above (Spencer et al. 1997), indicating temporal variability in plume gas abundance and/or composition. Finally, during a 6-month monitoring campaign in 1991–1992 Brown and Bouchez (1997) saw a large change in the Na emission intensity in the extended cloud, which they inferred was caused by a volcanic outburst, followed a short time later by an increase in torus S+ emission intensity, which they attributed to an increase in mass loading of the torus.

19.2.3 Recent Progress: Modeling

Although observations have begun to elucidate the nature of Io's atmosphere, they are still relatively infrequent and many gaps remain in our knowledge. Theoretical models provide complementary details that allow us to gain a more complete picture. Because comprehensive, self-consistent models are very difficult to construct, most attempts at modeling Io's atmosphere have focused on single aspects of the problem, namely vertical composition and density structure (Kumar 1982, Kumar and Hunten 1982, Kumar 1985, Summers 1985) or the horizontal distribution of surface pressure and its associated dynamics (Fanale et al. 1982, Matson and Nash 1983, Ingersoll et al. 1985, Ingersoll 1989, Moreno et al. 1991). Even though these simplified models had only moderate success in reproducing the Pioneer 10 ionospheric density profiles, they did indicate that Io's atmosphere must contain significant amounts of SO, O, S, and O. But the progress made in the last decade has now justified the development of more elaborate, two-dimensional models. We discuss here the latest efforts in the "single aspect" models, as well as the first attempts at "unified" models.

Modern Buffered Model

The early UV observations of Ballester et al. (1994) motivated Kerton et al. (1996) to reconsider the radiative equilibrium models of Fanale et al. (1982) because they gave SO2 abundances larger than observed. They rectified some of the oversimplifications in the treatment of radiative equilibrium by taking into account the latent heat of SO2 frost sublimation, the rotation rate of Io, thermal conduction and Io's internal heat flow, and the deposition of some solar energy below the surface (a component of the so-called "icy greenhouse" effect). Accounting for heat conduction shifts the maximum temperature slightly from the subsolar point toward the dusk terminator. The surface temperature and pressure gradients toward the periphery of Io's disk are much more gradual than for the radiative equilibrium case. These enhancements to the model resulted in reduced column abundances more consistent with the Ballester et al. (1994) results. An essential point is that these models don't assume a surface temperature dependence (and hence SO2 sublimation source) that is axisymmetric about the subsolar point. In the most extreme cases (the high conduction C/R/L – thermal conduction, rotation, latent heat – model, their Figure 6, and the subsurface greenhouse model, their Figure 8) the SO2 pressure near the poles is many orders of magnitude lower than near the terminators, which is qualitatively consistent with interpretations of the Lyman-α images (Feldman et al. 2000b, Strobel and Wolven 2001).

Volcanic Models

The more detailed plume composition information now available from the Spencer et al. (2000) and McGrath et al. (2000a) observations has enabled detailed thermochemical modeling of volcanic eruptions on Io. A series of papers by Zolotov and Fegley (Zolotov and Fegley 1998a,b, 1999, 2000, Fegley and Zolotov 2000) has presented results of ideal gas thermochemical equilibrium calculations to evaluate volcanic gas chemistry. Since some of the hot spot temperatures on Io range up to 1700 K, they argue that, by analogy with volcanic eruptions on Earth where volcanic gases erupted at temperatures &gt;900 K are high enough for thermochemical equilibrium, ionian gases may also chemically equilibrate during eruptions. They substantiated this by comparing the eruption times to the chemical lifetimes for volcanic gas chemistry. The eruption temperature, total pressure and bulk elemental composition of the volcanic gases are inputs to the modeling, and thermochemical equilibrium inside the volcano, and quenching in the vicinity of the volcanic vent are assumed. The composition of the erupted gas is then used to evaluate magma temperature and the oxidation state of the magma and Io's interior, to infer thermal and chemical conditions in the volcanic source region, and to calculate pressures in the vicinity of the vent. The earlier papers used compositional information from the Io plasma torus due to the lack of quantitative information about the plumes or atmosphere of Io. These models imply that the Pele plume gas last equilibrated at migmatic temperature and was not significantly altered in the eruption. The composition of the Pele plume indicates that Io is differentiated, and that metallic iron and free carbon are not abundant in bulk silicate on Io. Zolotov and Fegley (1998a) stress that SO is a natural product of thermodynamical equilibrium in erupted materials, and that the observed SO/SO2 mixing ratio can be fit for suitable combinations of gas pressure, temperature and a value of O/S &lt; 2 at the vent. They also predict S2O to be an important volcanic species. Also of
note is the detailed modeling of the temperature structure of volcanic plumes recently achieved by Zhang et al. (2003).

Radiative and Photochemical Models

The radiative and photochemical models described in this section involve one-dimensional calculations in which only the vertical thermal or density structure is modeled. In all cases the SO$_2$ density profile is assumed to be in hydrostatic equilibrium and not affected by photochemical and transport processes. Strobel et al. (1994) developed the first comprehensive model of Io's vertical thermal structure, extending and improving on the treatments by Kumar (1985) and Lellouch et al. (1992). They solved the time-dependent, one-dimensional heat balance equation with heat transport by diffusive and radiative processes, including solar heating in the UV and a detailed description of non-LTE cooling by SO$_2$ rotational and vibrational lines. Two cases were considered, a high-density atmosphere representative of the [smaller fractional coverage, higher column abundance] regime typified by the millimeter observations, and a low-density atmosphere intended to represent the [larger fractional coverage, lower column abundance] regime typified by the early disk-averaged UV observations. Their model predicts the existence of a mesopause in Io's atmosphere when the surface pressure exceeds ~10 nbar. None of the model atmospheres generated with only solar heating were hot enough to satisfy the hydrostatic interpretation of the millimeter data, or the bulk atmospheric temperature of 200–400 K derived from the UV data because the relevant temperature is the average temperature (~140 K) of the first scale height (~10 km). Two additional sources of heating were therefore explored: the heating associated with impacting thermal ions from the Io plasma torus as they sweep by Io's exosphere/upper atmosphere (Johnson and Matson 1989), and Joule heating, which is driven by the penetration of Jupiter's corotational electric field into Io's conducting ionosphere (see Figure 19.7). In agreement with the earlier results of Lellouch et al. (1992) they found that unless the plasma penetrates significantly below the exobase it would only elevate the exospheric temperature. In the limit of atmospheres with surface pressures in the range 0.1–1 nbar, Joule heating can in principle produce a warm atmosphere with average temperature greater than 200 K.

Making use of the thermal structure of Strobel et al. (1994), Summers and Strobel (1996) focused renewed effort on the photochemical modeling in order to gauge the sensitivity of the chemical structure to vertical transport rates, and to evaluate the possibility that O$_2$ and/or SO may be significant dayside constituents. Unlike the earlier photochemical models, they tested both low and high values of the eddy mixing rate. Minor molecular Na species were included in the calculations. The molecular and atomic Na escape rates of Wilson and Schneider (1994) and Smyth and Combi (1988), respectively, were used as constraints in the modeling, and they attempted to match the Pioneer 10 dayside ionospheric profile. Their results confirmed the prediction (Kumar 1985) that SO is an important atmospheric constituent. None of the cases considered could simultaneously both produce the large atomic and molecular Na escape rates, and provide a reasonable match to the Pioneer 10 ionospheric profile. They concluded that SO$_2$ photochemistry alone could not produce the O$_2$ column abundance suggested by Kumar and Hunten (1982), but that SO could potentially become a dominant background gas globally since it might not condense during Io night.

Moses et al. (2002a,b) have revisited the one-dimensional aeronomic models in order to address how active volcanism might affect the standard picture of photochemistry on Io. A variety of sodium, potassium, and chlorine-bearing volatiles are included, in addition to sulfur and oxygen species. Unlike the previous models, they assume that a Pele-type volcanic source continuously supplies gas at the

Figure 19.6. The SO$_2$ mm emission line area (left) and shift from rest frequency (right), which illustrate the variability of the SO$_2$ atmosphere discussed in the text.

Figure 19.7. The effects of solar (S), solar + plasma (S+P) and solar + plasma + Joule (S+P+J) heating (H) on the thermal structure of Io's atmosphere (Figure 12 from Strobel et al. 1994). Panel a: $T = 130$ K and $P = 130$ nbar; panel b: $T = 120$ K and $P = 3.5$ nbar at the surface.
surface. The Zolotov and Fegley type thermochemical equilibrium calculations described above are used to help constrain the composition and physical properties of the exsolved volcanic gases. The effects of photolysis, chemical kinetics, condensation, and vertical eddy and molecular diffusion are then tracked to determine the subsequent evolution of the gas. Moses et al. (2002a) focuses on sulfur and oxygen species. As might be expected, if S₂ is a common volcanic gas, the sulfur species (S, S₂, S₃, S₄, SO, and S₂O) are enhanced relative to the oxygen species (O and O₂) in their Pele-type volcanic models as compared with frost sublimation models. The indication of a higher SO/SO₂ ratio in the Pele plume (McGrath et al. 2000a) may reflect the importance of volcanic SO rather than low eddy diffusion coefficients or low SO surface “sticking” probabilities. If this ratio is affected by volcanic activity, it could be temporally and/or spatially variable to the extent that the S/O ratio in the neutral clouds and plasma torus could be correlated with volcanic activity.

Moses et al. (2002b) concentrate on alkali and chlorine species, predicting that NaCl, Na, Cl, KCl, and K are the dominant alkali and chlorine species generated from Pele-type eruptions. Although they test the sensitivity of their results to different assumptions about the gas composition, these five species dominate for a wide range of conditions. Other sodium and chlorine molecules are minor constituents of the atmosphere because of their low volcanic source rates and efficient photochemical destruction. NaCl is not recycled efficiently, as its loss rate by photolysis greatly exceeds the volcanic production rate. Unless a surface sputtered source is important it will be rapidly depleted from the atmosphere, and therefore Moses et al. (2002b) conclude with confidence that it should be more readily apparent during periods of higher volcanic activity. The observed NaCl mixing ratio (Lellouch et al. 2003) in volcanic gases is somewhat lower than expected from these thermochemical models because of the large abundances assumed for Na and Cl, which were based on extrapolations of torus Cl ion abundances. Estimates of the escape flux based on the NaCl volcanic rates inferred by Lellouch et al. (2003) and the Moses et al. (2002b) prediction for the fate of the NaCl ion, suggest that most of it escapes in the form of atomic Na and Cl at a rate consistent with the supply rates to the torus, as inferred from neutral cloud and torus luminosities.

"Unified" Models

Although models have yet to capture the full complexity of Io's atmosphere, the first steps have now been taken to combine, to some degree, the vertical and horizontal calculations in a series of papers by Wong and co-workers (Wong and Johnson 1995, 1996a, Wong and Smyth 2000). In contrast to the previous vertically averaged or static columns, they used a fluid dynamics model to perform two-dimensional calculations assuming a sublimation-driven SO₂ atmosphere with axial symmetry about the subsolar point. The first paper focused primarily on illustrating the effect of plasma heating on the sublimation-driven flow, and estimating the exobase altitude and torus supply rate; plasma and solar UV heating were included, but Joule heating and chemistry were not. As in previous work (Lellouch et al. 1992, Strobel et al. 1994), Wong and Johnson (1995) found that plasma heating is most important near the exobase, and thus has important consequences for the overall dynamics of the atmosphere in the region which controls the loss of atoms and molecules. Wong and Johnson (1996a) extended the analysis using an improved fluid dynamics model that also included photochemistry and diffusion. This paper was concerned mainly with evaluating the suggestion that noncondensibles (O₂, and possibly SO) could accumulate and dominate the atmospheric dynamics, and the question of whether a nightside O₂ atmosphere could build up. O₂ is assumed to be noncondensible, while SO is in some cases assumed to be noncondensible and in others to be condensible. They find that the buildup of a nightside atmosphere of noncondensible photochemical products does not overwhelm the dayside atmospheric flow, but its does raise the overall atmospheric pressure and reduce the wind speed. Wong and Smyth (2000) considered high and low density SO₂ atmospheres at both both dawn and dusk elongation using an improved version of the multispecies hydrodynamic code. Assuming that O₂ and SO are both noncondensible, they find that gas-phase reactions among the noncondensibles can produce a substantial amount of SO₂ in the nightside atmosphere. They consider four cases (high and low density SO₂ atmospheres at both eastern and western elongation) and find for all cases considered that an exobase above the surface occurs globally (see Figure 19.8). The dayside SO/SO₂ mixing ratio is ~3-7%, consistent with the SO measurements described above. The SO₂ subsolar column density in their models is always higher than the comparable hydrostatic sublimation atmosphere case, showing that the abundance of a sublimation atmosphere is dynamically controlled.

19.2.4 Synthesis

Our understanding of Io's atmosphere has improved radically in the nearly 30 years since its first detection. The current state of knowledge about it is summarized in Table 19.2. Since the detection of Na and the earliest models that assumed compositions of N₂, NH₃, CH₄, and Ne, nine species have now been directly detected (SO₂, S₂, SO, S, O, Na, K, NaCl, CI) and via theoretical modeling others (especially O₂) are strongly suspected. From models of the SO₂ atmosphere that ranged over at least six orders of magnitude in surface pressure, the plethora of new observations obtained since 1990 reach a number of reassuringly consistent conclusions, the firmest being that Io's SO₂ atmosphere is tenuous but collisionally thick, and is temporally variable but permanently detectable on both its leading and trailing dayside hemispheres. This readily excludes the purely subsurface cold trap and purely sputtered models. Further, the spatially resolved spectra provide evidence for a decrease in gas pressure with increasing latitude, and at least modest (factor of 2-5) density enhancements over active plumes. This conclusion is supported by the structure seen in the Lyman-α images and the most recent UV spectroscopy, and is in fact not contradicted by any observation. The decrease in SO₂ abundance with latitude is consistent with Io's surface temperature being lower at the poles than at the terminators (as predicted by the most recent radiative equilibrium models) and/or with volcanoes preferentially located at low latitudes. As with SO₂ frost (Douté et al. 2001), SO₂ gas is ubiquitous on Io, except perhaps at the poles.
Figure 19.8. The variation of column density with solar zenith angle for various constituents of Io's atmosphere in the 2-D, axisymmetric models of Wong and Smyth (2000) assuming SO is noncondensible. Left-hand figure is for western elongation, right-hand figure is for eastern elongation. (From Wong and Smyth 2000, Figure 5).

Table 19.2. Summary of Io atmospheric species.

<table>
<thead>
<tr>
<th>Species</th>
<th>Io Abundance*</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>SO₂</td>
<td>(<del>(1-10) \times 10^{16}) in (\pm (30-45)°) latitude band (</del>(2-10^?) \times) higher in active volcanoes</td>
<td>Synthesis of all observations; see Sections 19.2.2, 19.2.3, and 19.2.4; McGrath et al. 2000a; Spencer et al. 2000; Spencer et al. 2002; Jessup et al. 2004</td>
</tr>
<tr>
<td>S₂</td>
<td>(1 \times 10^{16}), Pele plume (t), SO₂/S₂ (~(3-12))</td>
<td>Spencer et al. 2000</td>
</tr>
<tr>
<td>SO</td>
<td>(~(0.03-0.1) \times SO₂)</td>
<td>Lellouch 1996</td>
</tr>
<tr>
<td>NaCl</td>
<td>((0.003-0.013) \times SO₂), active volcanoes</td>
<td>Lellouch et al. 2003</td>
</tr>
<tr>
<td>S</td>
<td>(3.6 \times 10^{12} &lt; N_S &lt; 1.3 \times 10^{14} (t)) (~9 \times 10^{12}) at (2 R_{Io} (t) = 0.1 \times O)</td>
<td>Feaga et al. 2002 (upper limit revised up; see text) Wolven et al. 2001</td>
</tr>
<tr>
<td>O</td>
<td>(~(4-7) \times 10^{13}), disk average (~1 \times 10^{14}) at (2 R_{Io} (t) = 11 \times S)</td>
<td>Ballester 1989 Wolven et al. 2001</td>
</tr>
<tr>
<td>Na</td>
<td>(4 \times 10^{12}), disk average</td>
<td>Bouchez et al. 2000 [see also Burger et al. 2001, Retherford 2002]</td>
</tr>
<tr>
<td>K</td>
<td>((1-10) \times 10^{6}); (Na/K = 10 \pm 5) at ((10-20) R_{Io})</td>
<td>Brown 2001</td>
</tr>
<tr>
<td>Cl</td>
<td>(~1 \times 10^{13}), disk average</td>
<td>Feaga et al. (2004)</td>
</tr>
<tr>
<td>H</td>
<td>(~2 \times 10^{12})</td>
<td>Strobel and Wolven 2001</td>
</tr>
<tr>
<td>CS₂</td>
<td>(&lt;2 \times 10^{14})</td>
<td>McGrath et al. 2000a; Spencer et al. 2000; Spencer et al. 2002</td>
</tr>
<tr>
<td>CO</td>
<td>(&lt;(3.6-6) \times 10^{17})</td>
<td>Lellouch et al. 1992</td>
</tr>
<tr>
<td>H₂S</td>
<td>(&lt;(0.7-1.2) \times 10^{16})</td>
<td>Lellouch et al. 1992</td>
</tr>
<tr>
<td>OCS, S₂O₂, ClO, CS, NaOH</td>
<td>Not detected (mm)</td>
<td>Lellouch et al. 1992</td>
</tr>
<tr>
<td>KCl</td>
<td>(&lt;1 \times NaCl)</td>
<td>Lellouch et al. 2003</td>
</tr>
</tbody>
</table>

* Numbers in vertical column density, cm⁻², unless otherwise noted; \((t) =\) tangential.
Most current evidence favors the [larger fractional coverage, smaller column abundance] regime as opposed to the [smaller fractional coverage, higher column abundance] regime. There is little compelling evidence for very large, localized enhancements except for the Voyager IRIS observation of Loki, and even that can be explained with a significantly lower column abundance than initially estimated (Lellouch et al. 1992). SO2 column densities inferred in the Pele plume from spatially resolved spectroscopy and imaging are in remarkable agreement at a value of \( \sim 3 \times 10^{16} \, \text{cm}^{-2} \). Thus, the emerging view is that of an SO2 atmosphere with a mean \( \sim (1-10) \times 10^{16} \, \text{cm}^{-2} \) column, covering typically 50–70% of Io’s dayside surface, mostly but not exclusively at low latitudes, with lateral variations encompassing a total factor of \( \sim 10–100 \).

However, as has been the case with Io’s atmosphere from the very beginning, it is currently not possible to reconcile all the observations. The “hydrostatic” interpretation of the millimeter observations of an atmosphere distributed in much denser and more localized centers, especially on the trailing side \( (f_b = 2.5–8 \%) \), is not consistent with the above picture. The extreme patchiness is relaxed in the “volcanic plume” interpretation, and larger surface coverages \( (f_b > 30 \%) \) are obtained when low gas temperatures \( (<200 \, \text{K}) \) are assumed. This requires the atmosphere to be dominated by direct output from ten to perhaps 20 volcanic centers. While it is difficult to assess the plausibility of such large numbers, it is worth emphasizing that we now have several direct indications that Io’s atmosphere is at least partly volcanic, including (i) the detection of SO2 in Pele’s plume at the limb (i.e., near the terminator, where the surface temperature is probably too low for a sublimation component); (ii) the detections of S2 and NaCl, both of which have short lifetimes in a hydrostatic atmosphere; (iii) the detection of the forbidden infrared SO band; and (iv) the temporal variability seen in the millimeter and Lyman-\( \alpha \) data and inferred from several other observations.

With typical sublimation rates of 1 mm yr\(^{-1} \) and an SO2 ice layer at least several centimeters thick (Schmitt et al. 1994), a sublimation atmosphere is expected to be relatively stable, whereas an atmosphere supplied by a large number of simultaneous volcanic plumes implies temporal variability. Volcanoes not only produce local atmospheres within the plumes, they also serve to maintain the frosts. All this suggests that the volcanic interpretation of the millimeter data should be preferred. A contrario, the widespread atmosphere seen in the HST spectra and Lyman-\( \alpha \) images may imply a sublimation component, although these data are inconsistent with an atmosphere that is azimuthally symmetric about the subsolar point. While plume-like models have not been used in fitting the UV data, it can be anticipated that they would lead to slightly higher SO2 columns, since these models include large column density variations on unresolved \((<100 \, \text{km}) \) scales. Preliminary assessment of the effect on the analysis of the Lyman-\( \alpha \) data suggests an \( \sim 50\% \) increase in the SO2 column, allowing further reconciliation with the millimeter data.

A discrepancy remains with the Galileo/UVS data, which depict Io’s atmosphere as consisting of two components, each of which is much denser than in the HST picture. If the spectral structure seen at 2200–2800 Å is due to SO2 gas, it is inevitable to find very high SO2 columns as the SO2 cross section in this range is less than \( \sim 1 \times 10^{-18} \, \text{cm}^2 \). It is also possible that the broadband structure is due to the surface reflectance of materials other than SO2 frost, which is the only surface component used in the models. SO2 patches with columns of \( 1 \times 10^{16} \) and even \( 4 \times 10^{17} \, \text{cm}^{-2} \) would be invisible (saturated out) at 2000–2200 Å and appear as black areas at Lyman-\( \alpha \). This may not be inconsistent if these dense patches are smaller than the \( \sim 200 \, \text{km} \) resolution limit of the Lyman-\( \alpha \) images. Conversely, the \( \lesssim 10^{17} \, \text{cm}^{-2} \) columns inferred from HST are undetectable in the UVS range.

The least well constrained parameter is the characteristic gas temperature. The Keck II observation of SO indicates that at least a fraction of the volcanic gas is hot \( (1000 \, \text{K}) \) – implying that the thermodynamics of eruptive plumes are more complex than a mere adiabatic expansion. In the millimeter, the hydrostatic interpretation of the best data gives \( T_{\text{atm}} = 200 \, \text{K} \) on the leading side and \( T_{\text{atm}} = 400 \, \text{K} \) on the trailing, but the preferred volcanic interpretation does not constrain the gas temperature. In the interpretation of the UV data, while many authors have simply assumed values for \( T_{\text{atm}} \), the temperatures inferred by Ballester et al. (1994), McGrath et al. (2000a), and Spencer et al. (2000) range from 110 to 500 K, with a general preference for 200–300 K. This problem is formidable, especially for a volcanic atmosphere which is expected to exhibit huge lateral temperature variations.

Despite considerable progress in the past 13 years, many questions remain, including a quantitative understanding of how variable the atmosphere is, whether such variability can be related to changes in the Io plasma torus, what the nightside atmosphere is like, and what happens to the atmosphere in eclipse. Further progress would be facilitated by new measurements of the SO2 frost reflectivity at 1200–3000 Å and reflectivities of other surface components are needed so that uncertainties in the interpretation of the UV data can be addressed. Although it is again formidable, true 3-D calculations, in which the assumption of axial symmetry about the subsolar point is relaxed, are now warranted by the observations. Electron excitation cross sections and rate coefficients are desperately needed to improve estimates of the atomic sulfur and chlorine densities in Io’s atmosphere.

19.3 EUROPA, GANYMEDE, AND CALLISTO

19.3.1 Early Work

Early work strongly suggested that the surfaces of Europa, Ganymede and Callisto are covered mainly with H2O frost (Kuiper 1957, Moroz 1961, Lewis 1971), which was first confirmed by the positive identification of H2O features by Pilcher et al. (1972). The subsequent report of an approximately 1 \( \mu \text{bar} \) atmosphere on Ganymede from a stellar occultation measurement (Carlson et al. 1973) motivated Yung and McElroy (1977) to develop a photochemical model of a sublimation-driven water ice atmosphere, which evolves into a stable molecular oxygen atmosphere by photolysis of H2O because the hydrogen preferentially escapes. In their model, nonthermal escape of O atoms balances the production of O2 to yield a surface pressure of \( \sim 1 \mu \text{bar} \) consistent with the Carlson et al. (1973) measurement. They reached several
conclusions: Ganymede should have an appreciable oxygen atmosphere as long as the partial pressure of atmospheric H₂O exceeds about 2 × 10⁻⁹ mbar, which occurs for surface temperatures ≥134 K; it should have a Lyman-α halo produced by resonance scattering of sunlight by the escaping hydrogen; and an O₂ atmosphere should have a significant ionosphere produced by photoionization of O and O₂ with a peak electron density of 2 × 10⁵ cm⁻³. Yung and McElroy also concluded that the higher albedo of Europa would inhibit sublimation, suppressing the formation of O₂; Callisto they deemed more promising because its surface temperature is higher. Kumar and Hunten (1982) pointed out that this model possessed an additional stable solution with a much lower surface pressure of ~10⁻⁶ μbar. Voyager 1 Ultraviolet Spectrometer stellar occultation measurements of Ganymede yielded an upper limit on the surface pressure of 10⁻⁵ μbar (Broadfoot et al. 1979), compatible only with the low pressure solution.

Prior to the Voyager 1 encounter with Jupiter, there was little appreciation of the importance of sputter generated atmospheres/exospheres. Significant advances were made in understanding sputtering processes by innovative laboratory studies (Lanzerotti et al. 1978, Brown et al. 1980, Johnson et al. 1983) and in application of these results to planetary systems (Johnson et al. 1982, Wolff and Mendis 1983). Based on laboratory data, Lanzerotti et al. (1978) suggested that bombardment of the satellite surfaces by the jovian plasma leads to an erosion rate on Ganymede that could support the H₂O partial pressure used by Yung and McElroy (1977). The rates would be much larger at Europa, but much smaller at Callisto. Subsequent laboratory data showed that O₂ is directly produced in and ejected from ice (Brown et al. 1980), a process referred to as radiolysis (see Chapter 20). Using these data, Johnson et al. (1982) estimated that O₂ sputtered from water ice on Europa could yield a bound atmosphere with a column density N₀₂ ~ 2–3 × 10¹⁵ cm⁻². Since O₂ does not stick efficiently at these temperatures and does not escape efficiently, the atmosphere is dominated by O₂, even though the sputtered flux of H₂O molecules is larger than that of O₂.

Wolff and Mendis (1983) assumed the basic validity of the Yung and McElroy (1977) model with water ice sublimation as the source of icy Galilean satellite atmospheres and estimated O₂ atmospheres with surface densities in the range of 10¹⁰–10¹⁴ cm⁻³, even though the Voyager UVS observations excluded such high densities on Ganymede and Callisto (Broadfoot et al. 1979, 1981) as pointed out by Evistat et al. (1985). For Europa, assuming a sticking coefficient of 1.5 × 10⁻³ for O₂, Evistat et al. (1985) calculated that the exobase is on Europa's surface, and they estimated that the "atmosphere" is a sputtered O₂ exosphere with column density ~1 × 10¹⁴ cm⁻².

19.3.2 Europa

Oxygen Atmosphere and Ionosphere

The first detection of Europa's atmosphere was accomplished using HST/GHRS by Hall et al. (1995), who discovered the semi-forbidden OI (³S⁻²P)1356 Å and optically allowed OI (³S⁻²P)1304 Å multiplets in emission (Figure 19.9). The observed intensity ratio, I(1356)/I(1304), was ~1.9:1. They interpreted this as evidence for electron impact dissociative excitation of O₂ as the dominant excitation mechanism. The other plausible sources of the observed OI multiplet, solar resonance fluorescence scattering by O atoms and reflection of solar UV light from the surface, produce negligibly small intensities. The inference that Europa's atmosphere is O₂ and not O is based on the I(1356)/I(1304) ratio. Noren et al. (2001) have provided accurate electron impact cross sections for these dissociative excitations. For a Maxwellian distribution of electrons over a broad energy range this intensity ratio is 2. By contrast, incorporating the OI 1304 Å cross section of Doering and Yang (2001), the I(1356)/I(1304) ratio for O atoms has a broad maximum of 0.35 at 4 eV.

The absolute intensities imply a molecular oxygen atmosphere with column density 1.5 ± 0.5 × 10¹⁵ cm⁻² (P₀ = 2.2 ± 0.7 × 10⁻⁹ μbar) on the trailing hemisphere of Europa, which is consistent with the early estimate of a bound atmosphere by Johnson et al. (1982) and the low pressure solution discussed by Kumar and Hunten (1982). In deriving the O₂ column density, Hall et al. (1995) assumed: the spatial distribution of Europa's atmosphere is confined to the geometric cross section of the observed hemisphere (i.e., the scale height of the atmosphere is significantly smaller than the radius of Europa); a negligible contribution to the observed flux is emitted from above the tangential limb along...
the terminator; the Io plasma torus electrons responsible for exciting the observed emissions interact with the atmosphere without energy degradation; and no electrodynamic, sub-Alfvénic interactions such as observed by Voyager at Io (e.g., Ness et al. 1979, Neubauer 1980) were considered. The oxygen atmosphere has been confirmed with additional HST/GHRS observations of both the trailing and leading hemispheres by Hall et al. (1998) (Figure 19.9). With a finite scale height and emission above the limb included, the inferred molecular oxygen column densities are in the range \( \sim (2.4-14) \times 10^{14} \text{ cm}^{-2} \).

Ip (1996) published an exospheric model for the \( \text{O}_2 \) atmosphere on Europa that included deflection of the incident plasma. He assumed that the atmosphere is created predominantly by magnetospheric thermal ion sputtering of \( \text{O}_2 \) from water ice. However, his calculated exospheric column density failed by more than a factor of 1000 to account for the inferred \( \text{O}_2 \) column density. He therefore invoked additional surface sputtering by newly ionized \( \text{O}_2 \) molecules accelerated by the corotational electric field and convected back into Europa’s surface from which they had only recently been sputtered. According to Ip this “resputtering” mechanism is sufficient to raise the density of the gravitationally bound \( \text{O}_2 \) molecules past the threshold value that defines the transition from an exosphere to an atmosphere.

Based on the HST observations and the previous modeling attempts, Saur et al. (1998) developed a plasma interaction model to account for the sources and sinks of the neutral atmosphere, and to describe the interaction of the jovian magnetosphere with the atmosphere and the formation of an ionosphere. They found that suprathermal torus ions plus a contribution from thermal ions could sputter \( \text{O}_2 \) from the water ice surface to create an atmosphere that is in mass balance with thermal ion stripping of the \( \text{O}_2 \) atmosphere by charge exchange and atmospheric sputtering. Using the sputtering rates of Shi et al. (1995), which are based on Voyager plasma measurements, an average \( \text{O}_2 \) column density of \( \sim (3-7) \times 10^{14} \text{ cm}^{-2} \) can be maintained. With this \( \text{O}_2 \) column density, they calculated intensities of the \( \text{O} \) 1356 and 1304 satellites in agreement with those observed by Hall et al. (1995). Europa’s “equilibrium” atmosphere is strongly influenced by the electrodynamic interaction. The magnetospheric ions that create the atmosphere experience a reduced effective geometric area for Europa as the plasma flow is shielded (Goertz 1980). As the atmospheric density increases, the ionospheric densities and conductivities are enhanced, leading to larger electric currents and reduced ionospheric electric fields. The resulting enhanced shielding reduces the sputtering rate. Stripping of the \( \text{O}_2 \) atmosphere by charge exchange and atmospheric sputtering increases initially at low atmospheric density and asymptotically approaches a constant rate consistent with a finite plasma power reservoir. Mass balance is achieved at the above column density range with atmospheric creation and removal rates of \( \sim 10^{27} \text{ s}^{-1} \) (see their Figure 2). Ionization of \( \text{O}_2 \) molecules followed by pickup and convection out of the atmosphere accounts for only 15\% of the atmospheric loss rate. Saur et al. (1998) showed that the Ip (1996) “resputtering” mechanism could contribute only \( \sim 0.005 \) of the total sputtering rate on Europa because ionospheric ions have low energy and sputtering yields.

Shematovich and Johnson (2001) have carried out a 1-D Monte Carlo simulation of Europa’s oxygen atmosphere. Representative molecules were ejected from the surface and followed between collisions, accounting for surface accommodation, dissociation and ionization by solar UV and magnetospheric electrons, and collisional ejection by the incident plasma. The non-thermal as well as thermalized \( \text{O}_2 \) were tracked, as were the energetic \( \text{O} \) atoms produced by dissociation. They used a recent set of cross sections for collisional dissociation and ejection of \( \text{O}_2 \) (Johnson et al. 2001). The atmospheric density, temperature and escape flux were calculated as a function of the surface source rate. They found that the atmosphere was not well approximated by the simple model in Saur et al. (1998), but they obtained a source rate required to match the oxygen observations that was comparable (Shematovich et al. 2004).

Using the Galileo plasma measurements, Cooper et al. (2001) carried out a comprehensive study of the effects of energetic ion and electron irradiation of the icy surfaces of the Galilean satellites. In their Table II they give an ion sputtering timescale for Europa of \( 6.1 \times 10^5 \text{ yr mm}^{-2} \), which is dominated by energetic oxygen and sulfur ions. When converted to a sputtered flux of \( 1.8 \times 10^9 \text{ cm}^{-2} \text{ s}^{-1} \) and multiplied by the surface area of Europa, one gets \( 5.6 \times 10^{23} \text{ s}^{-1} \) for the direct \( \text{H}_2\text{O} \) sputtering rate. Like the results in Shi et al. (1995) this is a lower bound to the sputtering rate which depends on temperature above 100 K. This result is much smaller than the Shi et al. (1995) \( \text{H}_2\text{O} \) sputtering rate of \( \sim 3 \times 10^{28} \text{ s}^{-1} \). The latter was based on Voyager measurements for which the suprathermal plasma was presumed to be predominantly heavy ions. However, the Galileo measurements clearly show that the ion energy flux to the surface is predominantly carried by the energetic protons and electrons (Cooper et al. 2001, Paranicas et al. 2001, 2002) which do not sputter as efficiently. The reduction of the \( \text{H}_2\text{O} \) yields due to surface porosity (Johnson 1989) is roughly accounted for in these estimates. Because the sputter-produced \( \text{O}_2 \) does not stick to neighboring grains there is no reduction for porosity, enhancing its relative source strength (see Chapter 20).

An ionosphere has also been detected on Europa. Kliore et al. (1997, 2001a) characterize results from the eight Galileo radio occultation measurements at Europa as five strong detections, two weak detections, and one non-detection of an ionosphere. They conclude on the basis of the various geometries for the angle, \( \Psi \), between the center of the trailing hemisphere (the sub-ram point) and the subsolar point that a necessary condition for detection of an ionosphere is that the trailing hemisphere must be partially solar illuminated, i.e., \( \Psi < 90^\circ \). The non-detection of an ionosphere in the E26 flyby entry occultation was at high latitude and in the wake region, where according to the plasma model of Saur et al. (1998) the ionosphere of Europa becomes detached from the satellite as it convects downstream (cf. their Fig. 7). Spherical symmetry, which is assumed in all analyses of ionospheric occultations, is then no longer applicable and the occultation geometry does not yield a Chapman factor column density enhancement of \( (2\pi R/H_p)^{0.5} \sim 10 \), where \( R \) is the satellite radius and \( H_p \) is the plasma scale height.

Saur et al. (1998) found that electron impact ionization can generate Europa’s ionosphere at the electron densities measured by Kliore et al. (1997, 2001a). The total ionization
rate ($\sim 10^{26} \text{cm}^{-2}\text{s}^{-1}$) is limited by the finite power supplied to the ionosphere by electron heat conduction along magnetospheric flux tubes that interact with Europa’s atmosphere. In their calculation they adopted typical Voyager magnetospheric plasma conditions at Europa: $n_e = 38 \text{ cm}^{-3}$ and $2 \text{ cm}^{-3}$ at $T_e = 20 \text{ eV}$ and $250 \text{ eV}$, respectively. The electron impact ionization rate is $1.9 \times 10^{-6} \text{s}^{-1}$, which may be compared to a solar maximum photoionization rate of $6 \times 10^{-8} \text{s}^{-1}$. It should be noted that the intrinsic time constants associated with these processes are 6 and 190 days, respectively, which should be compared with Europa’s 3.6-day orbital period. Thus the solar photoionization rate should be diurnally averaged (a factor of 0.5 for Europa’s optically thin atmosphere), whereas electron impact ionization depends mostly on ambient magnetospheric plasma densities. In order for photoionization to be competitive with electron impact ionization, the magnetospheric electron density would have to be $\lesssim 1 \text{ cm}^{-3}$. Thus in the case of Europa, it is difficult to understand why the existence of an ionosphere depends on solar illumination. For the Europa radio occultations observed in Galileo orbits E4 and E6, the ambient magnetospheric ion densities were $\sim 25$ and $15 \text{ cm}^{-3}$, respectively (Paterson et al. 1999), and the corresponding electron densities would be about 50% larger. Kurth et al. (2001) suggest that a typical torus electron density is $80 \text{ cm}^{-3}$ at the orbit of Europa.

Even more problematic is the lack of an ionospheric signature in the Paterson et al. (1999) and Gurnett et al. (1998) data for Europa’s wake region during the E4 flyby. Because ionospheric plasma should presumably be $E \times B$ convected downstream, Paterson et al. (1999) and Gurnett et al. (1998) should have detected plasma densities commensurate with the magnitude of the electron densities inferred by Kliore et al. (1997, 2001a). The plasma model of Saur et al. (1998) yielded reduced plasma velocities of $\sim 20 \text{ km s}^{-1}$ in the vicinity of Europa. If the peak density in the E4 wake data of Paterson et al. (1999) were interpreted as $O_2^+$, then the ion (and electron) density is at most $\sim 110 \text{ cm}^{-3}$ with speed $\sim 50 \text{ km/s}$. Similarly, Gurnett et al. (1998) found electron density enhancements of $50–100 \text{ electrons cm}^{-3}$ above the approximately 80 torus electrons $\text{cm}^{-3}$ at Europa’s orbit during the E4 flyby. Kurth et al. (2001) report plasma wave observations for the nine Galileo spacecraft close flybys of Europa. The maximum density enhancement observed on any flyby was on the E15 flyby when the electron density jumped from 200 to $400 \text{ cm}^{-3}$ as the spacecraft entered the wake. Because the E4 flyby was the closest to Europa in the wake, our analysis focuses on this flyby. The downstream distance from Europa is $\sim 2000 \text{ km}$ and the transit time is at most 100 s. An initial ionospheric parcel with $n_e = 10^5 \text{ cm}^{-3}$ will require $\sim 10^6 \text{s}$ to recombine dissociatively to a density of $1000 \text{ cm}^{-3}$, but almost an additional $10^6 \text{s}$ to decrease from 1000 to $100 \text{ cm}^{-3}$, as the recombination rate is proportional to $n_e^2$. To convect a distance of $2000 \text{ km}$ in $10^6$ and $10^5 \text{s}$ demands plasma speeds of 0.2 and 0.02 $\text{ km s}^{-1}$, respectively, completely out of the observational and modeling range of 20–100 $\text{ km s}^{-1}$.

Recent HST/STIS images in the OI multiplets (Figure 19.10) indicate a more complex pattern of emission than would be expected from plasma interaction with an optically thin atmosphere. The McGrath et al. (2000b) OI 1356 image displays the expected limb glow around the disk plus a much brighter region on the anti-jovian hemisphere. Shown for comparison is the calculated intensity pattern from the Saur et al. (1998) model which emphasizes the bright limb glow due to the long tangential path length above the limb. The STIS point spread function will degrade the sharpness of predicted limb glow to the more diffuse glow seen in the STIS image. It is difficult to understand why the brightest region is on the disk due to its reduced path length. Since Europa has a weak induced magnetic field (Kivelson et al. 1997), it is not capable of focusing jovian electrons or energetic ions with finite gyroradii which sputter molecules into localized regions. It would be much easier to understand the bright region if it were on the limb in an asymmetric atmosphere with larger abundance on the anti-jovian side and the same physics that produces Io’s equatorial bright spots, where the brighter one is on the anti-jovian side. The observation of localized O emission may suggest that the surface is not icy everywhere, as assumed in the models described above, but rather that the composition varies considerably with longitude.

In summary, the HST observations of Hall et al. (1995, 1998) and the ionospheric radio occultations of Kliore et al. (1997, 2001a) lead to a consistent description of Europa’s atmosphere with the interpretive aid of a model that includes plasma deflection like that of Saur et al. (1998). It should be emphasized that the ratio of the ion drag force due to plasma flow through Europa’s atmosphere to the gravity force $(\sigma \times v_i) n_i v_i / g$ is large ($\sim 4$), for $10^4$ electron $\text{cm}^{-3}$ and $v_i \sim 50 \text{ km s}^{-1}$, where a typical collision rate for $O_2^+$ with $O_2$ of $(\sigma \times v_i)$ is $\sim 10^{-8} \text{ cm}^3 \text{s}^{-1}$. Consequently, ionospheric plasma should be convected downstream, yet plasma observations by Paterson et al. (1999) and Gurnett et al. (2001) have not detected an Io-like ionospheric wake.

**Alkalis at Europa**

Both Na and K have been observed at Europa (Brown and Hill 1996, Brown 2001) and they occur in a ratio both very different from that at Io, and from the meteoritic or solar abundance ratios (Brown 2001, Johnson et al. 2002). The description of the sodium cloud at Europa is simpler than it is at Io because Europa has a much thinner atmosphere, where collisions are few and atoms ejected from the surface can directly escape. The morphology of the observed sodium cloud can therefore be used to constrain the ejecta velocity distributions and to directly estimate the surface source rate.
and the alkali concentration in the surface (Johnson 2000). Leblanc et al. (2002) have produced a model of the sodium cloud at Europa (Figure 19.11) which is consistent with the observations. The model used to create Figure 19.11 includes the gravity of Jupiter, giving a banana-shaped cloud like that at Io (Smyth and Combi 1997), and solar radiation pressure, which stretches the cloud in the direction away from the Sun. Since most of the ejected Na and K atoms return to the surface, the alkalis are efficiently redistributed across Europa's surface from their initial source region. By fitting to Brown's observations, a source rate is obtained, as well as the velocity distribution for sodium ejected from the surface. Quite remarkably, the velocity distribution obtained is very similar to that recently measured by Yakshinskiy and Madey (1998, 2001) for alkalis ejected from ice by electronic sputtering. Using this model the Na/K ratio close to Europa has been estimated (Johnson et al. 2002). The average surface concentration of Na is ~0.5% and the required source of sodium is larger than that supplied by micrometeorites or by plasma ion implantation from Io. Since the Na/K ratio also differs from that at Io (Brown 2001), a subsurface source of alkalis is suggested. The Na/K ratio at Europa, though somewhat higher than predicted by models of a subsurface ocean, is not inconsistent with fractionation during transport to the surface (Zolotov and Shock 2001). The observed alkalis may therefore be the first direct detection of material from Europa's subsurface ocean.

19.3.3 Ganymede

Hall et al. (1998) also discovered an atmosphere on Ganymede via HST/GHRS observations of the same oxygen multiplets observed on Europa (Figure 19.9). As for Europa, the intensity ratio OI 1356/OI 1304 of ~1.3 implies a predominantly O₂ atmosphere. They estimated molecular oxygen column densities in the range of ~(1-10) × 10^{14} cm⁻². The OI 1356 multiplet structure exhibited a doubly-peaked profile that implied non-uniform spatial emission consistent with two distinct polar cap regions. These regions were thought to correspond to the loci of open field lines of Ganymede's intrinsic magnetic field (Kivelson et al. 1996).

The spatial distribution of the oxygen emission was clarified by the HST/STIS observations of Feldman et al. (2000a) (Figure 19.12), which indicate that the brightest Ganymede OI emissions are auroral in a manner analogous to the Earth's highly variable auroral oval regions. They correspond to regions where electrons trapped inside Ganymede's magnetosphere can be accelerated down field lines into the atmosphere to generate high conductivity and luminosity.

Based on excitation rates for Maxwellian electrons in the range Te = 10–20 eV, we estimate the OI red line at 6300 Å should be about 10 times as bright as OI 1356 Å. Brown and Bouché (1999) used the Keck telescope to detect the OI red lines (6300 Å, 6364 Å) when Ganymede was in eclipse. The estimated red line intensity of 1–2 K R is consistent with STIS inferred intensities for OI 1356 Å. They observed two “spots” near opposite limbs at low latitudes, however, the spatial resolution is only ~1 Ganymede radius, making it difficult to directly compare the morphology with the STIS observations. Although polar spots can be ruled out, the observed spots could be equatorial atmospheric limb emission confined to regions with closed magnetic field lines or alternatively, the mid-latitude “auroral oval” emission patterns such as those seen in the HST/STIS observations.

Barth et al. (1997) measured HI Lyman-α emission from Ganymede's atmosphere of 0.56 K R with the Galileo Ultraviolet Spectrometer. From the radial intensity profile, they deduced an H atom density of 1.5 × 10^{14} cm⁻³ at the surface with exospheric scale height equal to 1 Ganymede radius (= 2634 km) and temperature of 450 K. Feldman et al. (2000a) also detected Lyman-α emission in their HST/STIS observations. Their radial intensity profile, and hence density profile, was in excellent agreement with the Barth et al. (1997) data.

Unlike Europa, Na has not been detected at Ganymede. Long slit high resolution spectra obtained by Brown (1997) provide only an upper limit of 10⁶ atoms/cm² between 7800 and 15 600 km from the satellite surface, which is 13 times smaller than the Na density at the same distance from the surface of Europa. This is roughly consistent with the lower sputtering rates (e.g., Cooper et al. 2001) and higher es-

Figure 19.11. The Na intensity observed in N–S (left) and E–W scans of Europa compared with the model (solid line) of Leblanc et al. (2002) described in the text. Crosses include a background subtraction, open circles do not. From Figure 8 of Leblanc et al. (2002).
cape energies for Ganymede, but also likely means that the concentration of Na on the surface is much smaller on Ganymede than on Europa.

In sharp contrast to Europa, Kliore et al. (2001b) reported only one strong detection, two weak detections, and 5 non-detections of an ionosphere via Galileo radio occultation measurements. Most of these occultation observations occurred at low latitudes where Ganymede’s magnetic field dominates and the field lines are closed. Two occultations occurred at latitudes of 47–49° in the vicinity of the separatrix region, but with an angle $\Psi \sim 145°$. The strong detection and one of the weak detections occurred at a latitude of 20° and $\Psi = 48°$, but no ionosphere was detected at latitude 7° and $\Psi = 68°$. The other weak detection occurred at latitude $-25°$ and $\Psi = 94°$ for the entry occultation, but not for the exit occultation. Thus the requirement of $\Psi < 90°$ (which corresponds to partial solar illumination) applies to the strong detection but is marginally violated by the weak detection for $\Psi = 94°$.

Ganymede, with its intrinsic magnetic field, presents a bigger challenge to constraining or inferring the O$_2$ column abundance. A distinct polar limb glow would provide the best inference of O$_2$ column densities. Because the intensity ratio of the OI multiplets is consistent with electron impact on O$_2$, the inferred O$_2$ abundance depends on the electron density and energy distribution function. The broad polar caps on Ganymede (latitude >45°) are open to jovian plasma. The plasma densities and distribution functions have been measured by both Voyager and Galileo plasma instruments (Scudder et al. 1981, Paterson et al. 1999). Voyager measurements in the plasma sheet found enhanced densities ($\sim 5-20$ cm$^{-3}$) with $T_e \sim 20$ eV, and a suprathermal tail at 2 keV and 0.1 times the core density (Scudder et al. 1981). This enhanced electron population could support OI 1356 Å limb glow of 10–40 R in an atmosphere constrained by an upper limit deduced from a Voyager UV stellar occultation (Broadfoot et al. 1981). However, the STIS observations display polar limb glow in the range 50–100 R, suggesting that more than plasma sheet electrons are involved in the excitation process.

The latter point is more evident when an explanation is sought for the hot spots of intense auroral emission seen in the HST/STIS observations (see Figure 19.12 and Figure 3 of Feldman et al. 2000a). Eviatar et al. (2001) addressed this question by asking how many electrons at what temperature are needed to generate one of the observed 300 R bright spots. They found a threshold electron number density of 310 cm$^{-3}$ at $T_e \sim 100-200$ eV, more than an order of magnitude larger than the densities and temperatures measured in the plasma sheet. The hot spots could be produced by higher density, lower temperature electrons (e.g., $n_e = 2500$ cm$^{-3}$ and $T_e = 8$ eV). All of these values pertain to an atmosphere ($N_0 = 2.5 \times 10^{14}$ cm$^{-2}$) that satisfies the Voyager UV stellar occultation upper limit. Eviatar et al. argued that in the absence of any measured electron population to produce the auroral hot spots, local acceleration of plasma is required. They identified stochastic acceleration by electrostatic waves and/or magnetic field-aligned electric fields with associated Birkeland currents as possible mechanisms. The hot spots are then regions of high electron density ($\sim 10^5$ cm$^{-3}$), high conductance ($\lesssim 100$ mho), and high “auroral electrojet” currents, all features of the Earth’s auroral oval. The regions of high electron density would not be detectable by radio occultation measurements because of their limited spatial dimensions. One other potential solution to understanding the high HST intensities is to argue that the Voyager UV stellar occultation measurements are not applicable to the Galileo/HST epoch and atmospheric column densities are perhaps an order of magnitude larger, $\sim 3 \times 10^{15}$ cm$^{-2}$, which is the column density that stops 20 eV electrons.

In summary, our limited information prevents a definitive inference of the average O$_2$ column density on Ganymede. It is improbable that it is less than $10^{14}$ cm$^{-2}$ to account for the HST intensities and debatable that it exceeds $3 \times 10^{15}$ cm$^{-2}$ in the polar cap regions.

### 19.3.4 Callisto

Much less is known about the atmosphere of the most distant Galilean satellite, Callisto. However, several recent results bear on this question. The first is the report by Carlson (1999) of a Galileo NIMS observation of limb emission from the 4.26 μm ν$_2$ fundamental stretching band of CO$_2$ up to an altitude of 100 km above the surface of Callisto (Figure 19.13). The geometry of the observation was such that it mapped a region tangential to the limb very near the equator and near noon, in the wake region of the plasma flow relative to Callisto. The observed emission was attributed to resonance fluorescence of sunlight by atmospheric CO$_2$. Modeling the data assuming an isothermal atmosphere, the best fit was obtained for a temperature of 150 K (close to the measured noon surface temperature of Hanel et al. 1979, Spencer 1987) and surface pressure of 7.5 pbar. The scale height of the assumed exponential atmosphere is 23 km, and the surface number density is $\sim 4 \times 10^9$ cm$^{-3}$, implying a vertical CO$_2$ column abundance of $\sim 8 \times 10^{14}$ cm$^{-2}$ with estimated errors of about ±60%. This is comparable to the...
O₂ column densities on Europa and Ganymede inferred by Hall et al. (1998).

McCord et al. (1997) had previously noted the presence of absorption features in infrared spectra of Ganymede and Callisto that included the 4.26 μm band, which was attributed to CO₂ trapped in fine water ice. This feature appears to be considerably stronger in the detailed mapping spectra of Callisto (Hibbitts et al. 2001). This does not necessarily imply a higher abundance of CO₂ on Callisto but may reflect a different nature of the host water ice. However, STIS observations of Ganymede (Feldman et al. 2000a) failed to disclose the presence of carbon atoms or any carbon-bearing molecules.

Second, Gurnett et al. (2000) have reported Galileo plasma wave measurements that imply the presence of electrons with a density almost a thousand times higher than the expected jovian magnetospheric electron density at the orbit of Callisto. This density is comparable with that inferred from similar measurements made in the vicinity of Ganymede. Furthermore, the measurements suggest that the maximum density occurs in the geometric wake of the co-rotational plasma flow, but not far downstream in an Io-like tail, so one would expect to see a strong trailing/leading hemispherical asymmetry in the atmospheric emissions excited by the plasma electrons. These few measurements could also be interpreted to imply that solar illumination is essential for the presence of the highest plasma densities.

Third, Galileo magnetometer measurements (Khurana et al. 1998, Kivelson et al. 1999) show that Callisto, like Europa, has large-scale magnetic field perturbations characteristic of an induced magnetic dipole field as suggested by Neubauer (1998). For Callisto the perturbation amplitude is equal to the background jovian magnetic field, ~35 nT.

Fourth, a search for UV emissions diagnostic of the presence of O₂, CO₂, and/or CO atmospheres by Strobel et al. (2002) yielded only upper limits of 5 × 10⁻⁴ photons s⁻¹ or 15 R for a uniform disk the diameter of Callisto for emissions of Ω λ 1304 Å, Ω λ 1356 Å, Ω λ 1361 Å, Ω λ 1335 Å and CO Fourth Positive bands. No useful upper limits on O₂, CO₂, and CO atmospheres, in comparison to the detected CO₂ atmosphere by Carlson (1999) and the inferred O₂ atmosphere by Kliore et al. (2002) were derived (see below), but respective upper limits of ~10⁻¹³ and 2.5 × 10⁻¹³ cm⁻² were placed on atomic carbon and oxygen abundances.

Finally, we note the Galileo radio occultation detection of ionospheric electrons with densities up to 2 × 10⁴ cm⁻³ by Kliore et al. (2001b), when the necessary geometric configurations of solar illumination and ram angle were met during flybys C20, C22, and C23, when Ψ was ~2°. These detections imply the presence of a thin, non-uniform atmosphere, which Kliore et al. (2002) argue should be O₂, by analogy with Europa, with column density ~3 × 10¹⁴ cm⁻² and surface density ~2 × 10¹⁰ cm⁻³. These densities are comparable to the densities on Io and almost a factor of 100 times the densities on Europa and Ganymede.

During the Galileo flybys with occultations, only C22 yielded plasma wave emissions suitable for extraction of local electron densities (Gurnett et al. 2000), but unfortunately these in situ measurements were in the geometric wake on the dark hemisphere side. At closest approach, ~2300 km, no electrons were discernible (n_e < 0.2 cm⁻³), and further away the torus electron densities were at most 1 cm⁻³ and probably at least an order of magnitude lower. By comparison the ionospheric electron densities measured by radio occultation were ~15 000 cm⁻³ (entry), and ~8000 cm⁻³ (exit) (Kliore et al. 2001b), with no detectable wake extension as on Io.

The low torus electron densities imply photoionization is the dominant ionization mechanism in Callisto’s atmosphere and in this context Callisto’s ionospheric densities are extremely large. It is also surprising that with potentially large J × B forces the ionospheric plasma is not detectable as it is convected downstream, since even for molecular ions, there is insufficient time for the ionospheric plasma to recombine to densities below the detection threshold of the Galileo plasma wave instrument. However, Jupiter’s magnetic field is very weak (~35 nT) at Callisto. With Carlson’s derived CO₂ atmosphere and the measured electron densities of Kliore et al. (2001b), Strobel et al. (2002) obtained peak Pedersen and Hall conductivities of ~0.02 and ~0.01 m⁻¹ at ~35 and 27 km, respectively. The corresponding peak Pedersen and Hall conductances are ~15 000, and ~7000 mhos. A denser atmosphere due to a larger O₂ component as Kliore et al. (2002) infer would not significantly alter these estimates, as the conductivities already maximize above the surface. These conductances are enormous in comparison to the Alfvén conductance, Σ_Α ~1.3 mhos and, regardless of the details of the magnetic field line topology close to Callisto, Σ_P and Σ_H are ~Σ_Α. Strobel et al. (2002) estimated that these huge ionospheric conductances result

Table 19.3. Summary of Europa, Ganymede, and Callisto atmospheric species.

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<thead>
<tr>
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<tr>
<td>O₂</td>
<td>(2.4–14) × 10¹⁴ 1</td>
<td>(1–10) × 10¹⁴ 1</td>
<td>3 × 10¹⁴ 6</td>
</tr>
<tr>
<td></td>
<td>(3–7) × 10¹⁴ 2</td>
<td></td>
<td>(inferred, ionosphere)</td>
</tr>
<tr>
<td>CO₂</td>
<td>–</td>
<td>–</td>
<td>8 × 10¹⁴ 7</td>
</tr>
<tr>
<td>O</td>
<td>&lt;0.1 × O₂ 1</td>
<td>–</td>
<td>&lt;10¹³ 8</td>
</tr>
<tr>
<td>Na</td>
<td>4 × 10⁸ at 5 RE 3</td>
<td>&lt;1 × 10⁸ at 3–6 RE 4</td>
<td>–</td>
</tr>
<tr>
<td>K</td>
<td>1.6 × 10⁸ at 5 RE 3</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>C</td>
<td>–</td>
<td>–</td>
<td>&lt;2.5 × 10¹³ 8</td>
</tr>
<tr>
<td>H</td>
<td>–</td>
<td>2.4 × 10¹² 5</td>
<td>–</td>
</tr>
</tbody>
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* Numbers in vertical column density, cm⁻², unless otherwise noted; (t) = tangential (1) Hall et al. 1998; (2) Saur et al. 1998; (3) Brown 2001; (4) Brown 1997; (5) Barth et al. 1997; Feldman et al. 2000a; (6) Kliore et al. 2002; (7) Carlson 1999; (8) Strobel et al. 2002.
in \( \sim 1.5 \times 10^5 \) A flowing through Callisto’s ionosphere, which shorts out the corotation electric field yielding a highly reduced ionospheric electric field, and severely retarded ionospheric convection, which increases the \( \mathbf{E} \times \mathbf{B} \) plasma transport time constant far in excess of the plasma recombination time constant. The implication is that the observed CO\(_2\) atmosphere can support the observed ionosphere as a static, gravitationally bound plasma impervious to Jupiter’s corotational electric field and the associated \( \mathbf{E} \times \mathbf{B} \) convection of the ionosphere from the dayside to the nightside and/or off Callisto. The inability to detect UV emission can be understood by this strong electrodynamic interaction with the jovian magnetosphere that reduces the net electron impact emission rate by a factor of \( \sim 1500 \).

In addition, Callisto’s ionospheric conductances satisfy the total conductance required by Zimmer et al. (2000) to explain the induced magnetic field inferred by Kivelson et al. (1999). However, Callisto’s ionosphere is transient, present only when the trailing (ram) hemisphere is partially solar illuminated (Kliore et al. 2002), and thus cannot be the sole explanation for its induced field. The key remaining problem is the large atmospheric densities (\( \sim 40 \) times the Carlson (1999) CO\(_2\) density) that Kliore et al. (2001b) require to explain the inferred ionosphere. With noon temperatures on Callisto of \( \sim 150 \) K, the vapor pressure of H\(_2\)O is \( \sim 3 \times 10^3 \) cm\(^{-3}\). The ionization rate of H\(_2\)O is \( \sim 1.2 \times 10^{-8} \) s\(^{-1}\). In photochemical equilibrium with an effective dissociative recombination rate at elevated electron temperatures of \( 1 \times 10^{-7} \) cm\(^{-3}\) s\(^{-1}\), the surface electron density would be close to \( 2 \times 10^4 \) cm\(^{-3}\) at the terminators the surface temperatures are much colder and a water atmosphere would presumably condense and leave photochemically produced O\(_2\) as the dominant component of the atmosphere. It might be tempting to argue that the terminal ion is Na\(^+\), but like Ganymede, a search for Na atoms at Callisto has given a null result, and no upper limit exists (M. E. Brown 2002, personal communication).

19.4 SUMMARY AND OUTSTANDING QUESTIONS

The Galilean satellites possess tenuous atmospheres with surface pressures ranging from nanobars (SO\(_2\) on Io and the O\(_2\) inferred to be present on Callisto) to picobars (O\(_2\) on Europa and Ganymede; CO\(_2\) on Callisto). A summary of the atmospheric species and their abundances for Io is given in Table 19.2, and for Europa, Ganymede, and Callisto in Table 19.3. Io’s atmosphere, with typical SO\(_2\) column densities \( \geq 10^{16} \) cm\(^{-2}\), is collisionally thick near the equator, but decreases in density toward the poles by more than two orders of magnitude to a collisionally thin value. The distributions of atmospheric species in the atmospheres of Europa, Ganymede, and Callisto are not presently well understood. Despite considerable progress in characterizing the Io atmosphere quantitatively in the past 20 years, and in detecting tenuous atmospheres at the other Galilean satellites, many questions remain.

- Is the predominant source of Io’s atmosphere sublimation or volcanoes?
- How variable is Io’s atmosphere (both temporally and spatially), and does it drive Io plasma torus variability? What happens to Io’s atmosphere at night, and during eclipse?
- What is the relative abundance of O vs. O\(_2\) in the Europa and Ganymede atmospheres and how are they distributed spatially?
- Why is Europa’s atomic oxygen emission (indicative of its O\(_2\) atmosphere) apparently non-uniform?
- Is the ultimate source of the alkalis at Europa a subsurface ocean?
- What happens to the ionospheric plasma convected downstream at Europa, and why was it not detected by Galileo measurements?
- What is the average O\(_2\) atmospheric density at Ganymede? Do the UV and visible oxygen emissions detected at Ganymede have a common (auroral) origin?
- What is the source of Callisto’s atmosphere, endogenic or exogenic? Is the major species inferred from the ionospheric measurements O\(_2\), as is presumed?

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