Table 2 Fits to Plumes 1 and 3 for the aerodynamic model

Plume no.	Observed i_0 (deg)	From (21), (22)* q_0^2/g (km)	z_0 (km)	From (18) r_m (km)	z_m (km)
1	30	550	29	449	65
3	45	137	7	112	17

Plume no.	From (19) r_m (km)	z_m (km)	From (23) h_p (km)	q_m (km s^{-1})	q_p (km s^{-1})
1	212	145	215	0.33	1.00
3	91	39	54	0.16	0.51

* Equation nos given in parentheses.

Comparison and observations

Solution of equations (21) and (22) yields the results in Table 2. The quantity q_m is the velocity just below the shock on the axis of the plume computed by application of the conservation of energy which yields:

$$q=[q_0^2-2g(z_0+z)]^{1/2} \tag{24}$$

The disagreement between the computed height of the plume from the measured r_m, r_p and the measured height probably occurs because the actual trajectories beyond the shock disk are lower than the ballistic trajectories used. This is also indicated by the outer bend in the filaments appearing in Fig. 1. The inner bend is, of course, imposed by the shock front.

The thermodynamic diagram for sulphur dioxide prepared by Smith *et al.*[5] and their discussion make it clear that we are in a regime in which both the solid and the vapour are present in roughly equal amounts. In that case passage through the shock front will take place at a ratio of specific heats, close to unity with a large density increase, significant sublimation of entrained crystals of sulphur dioxide and that the temperature immediately above the front will be much lower than that which would occur for a perfect gas. The downward flow would involve expansion into a volume only a few times larger and recompression on reaching the surface could produce another shock front in the shape of a ring with lateral escape, unless most of the gas immediately crystallises onto the surface. The observations do not show such a front so it is either very close to the surface or does not exist owing to rapid crystallisation onto the surface.

Comparison of *Plate 7b*, p. 789 with Fig. 1 strongly suggests that we are witnessing variable rates of injection of another substance (sulphur droplets?) into a steady gaseous flow (sulphur dioxide).

These results were made possible by all those who have worked on the Voyager Project. Many of these have been individually acknowledged in ref. 1. We thank C. C. Avis, G. W. Garneau, P. L. Jepsen, J. J. Lorre, J. A. Mosher, G. M. Yagi, J. T. Harwood, L. Pieri, and C. Reslock of JPL and also S. A. Collins as experiment representative at JPL. This paper reports one phase of research supported by NASA and JPL under contract NAS7-100 and of research supported by JPL at the Smithsonian Astrophysical Observatory under that NASA contract.

Received 19 July; accepted 3 August 1979.

1. Smith, B. A. *et al. Science* **204,** 951–972 (1979).
2. Strom, R. G., Terrile, R. J., Masursky, H. & Hansen, C. *Nature* **280,** 733–736 (1979).
3. Prandtl, L. *Essentials of Fluid Dynamics* (Hafner, New York, 1952).
4. Oswatitsch, Kl. *Gas Dynamics* (Transl. Kuerti, G.) (Academic Press, New York, 1956).
5. Smith, B. A., Shoemaker, E. M., Keiffer, S. W. & Cook, A. F. *Nature* **280,** 738–743 (1979).

Volcanic resurfacing rates and implications for volatiles on Io

T. V. Johnson*, A. F. Cook II†, C. Sagan‡ & L. A. Soderblom§

* Jet Propulsion Laboratory, Pasadena, California 91103
† Smithsonian Astrophysical Observatory, Cambridge, Massachusetts 02138
‡ Cornell University, Ithaca, New York 14853
§ US Geological Survey, Flagstaff, Arizona 86001

Resurfacing rates and surface ages on Io are estimated, together with the material ejection and deposition rates of the active volcanic plumes.

THE discovery of current volcanism on Io[1,2] raises important new questions about the history of volatile material on Io, the rates of material deposition, the age of the currently observed surface, the loss rates of volatiles, and the ultimate fate of volatiles. This article presents an initial comparison of resurfacing rates based on two separate observations and implications for supply of material to the jovian torus. In estimating the resurfacing rates and surface ages we first examine the lack of recognisable impact craters down to the limiting resolution of about 600 m. Second, we estimate from imaging data the material ejection and deposition rates of the active volcanic plumes.

Removal of impact craters

To estimate the age of Io's landforms and hence their removal rate based on the absence of impact craters we must assume some flux of impacting bodies. Although such an estimate has large error bounds we are asking order-of-magnitude questions: is the resurfacing rate mm per yr, mm per millenium, or mm per aeon? Are the surface landforms hundreds of thousands or hundreds of millions of years old? Two methods have been used to estimate current relative impact rates for the terrestrial planets—Mercury, the Moon and Mars. The first models the distribution and impact rates from telescopically observed asteroids, comets and meteor showers[3,4]. The second compares the variations in the number of small impact craters (a few kilometres diameter) formed on smooth plains on these bodies following the extremely rapid decline in impact flux ~4,000 Myr ago[5]. Both these techniques suggest that the impact fluxes in the Solar System over a heliocentric range of 0.4 to ~1.5 AU have been roughly uniform, within a factor of two, during the past 10^9 years or so. The extrapolation to the jovian system suggests that fluxes there cannot be different by more than an order of magnitude from those in the inner Solar System without violating the apparently uniform character of the inner fluxes. Hence for our order of magnitude estimates we will use fluxes quoted for the Moon and Mars[5] realising that they might be in error by as much as a factor of 10 due to undiscovered populations, Jupiter gravity effects, and so on.

Figure 1 shows the three steps used to estimate that resurfacing rate which will just erase the expected impact craters.

0028–0836/79/350746—05$01.00

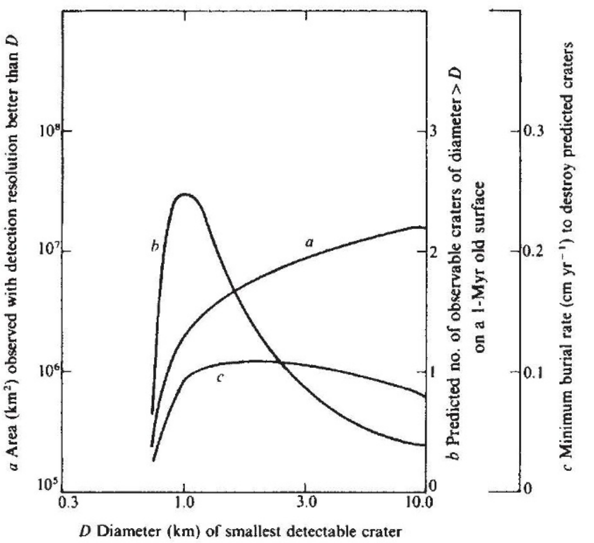

Fig. 1 Estimate of resurfacing rate to destroy impact craters. Shown as a function of crater diameter are: *a*, the areal coverage (in km^2) of Io obtained by Voyager 1 on which a crater of diameter D or larger could be detected; *b*, the number of craters larger than D that would have been seen in the Voyager 1 picture collection if Io's surface were 1 Myr old, assuming a lunar cratering size frequency spectrum and rate; *c*, the minimum rate of burial (or erosion) required to remove the craters so that none are now visible.

Different estimates can be made for different crater sizes. For instance, although larger craters are less frequent, and on the average older, they may take longer to destroy. Furthermore although smaller craters are more frequent they require higher resolution pictures to be recognised. Since in general we have less coverage at the higher resolutions, there is a higher probability of seeing a larger crater than a small one in many cases. Figure 1 shows that for the Voyager 1 pictures the probability of observing a 1 km crater is higher than that for a 600 m crater, although the latter would be far more abundant on the surface. The minimum resurfacing values in Fig. 1 are calculated assuming that the average crater diameter larger than D (the diameter of Io) has a diameter $\sim D$, that the craters have a depths of $1/5D$ and that burial by a blanket twice their depth is required to obscure them. The best estimates come from observations of 1–5 km craters, which should be statistically common and easily observed. Our value of the required resurfacing rate 10^{-1} cm yr^{-1} and should be reliable within a factor of 10. Hence our lower limit is 10^{-2} cm yr^{-1}. Also, resurfacing can be performed by a variety of processes including deposition by volcanic surface flows[6], volcanic plumes[7] and surface erosion[8]. Do the observed active plumes, which may be the most active of these processes, provide material at a rate consistent with these burial rates?

Depositional rates of volcanic plume material

To estimate the resurfacing rate from observations of current volcanic activity we start with the dimensions of plumes of volcanic ejecta[7] (r_p is the radius of plume; h_p the height of plume: see Table 1 in ref. 7). The visibility of the plumes in Voyager 1 images implies an optical depth of the order of at least unity in the denser parts of the plume along an optical path of scale $\sqrt{2}r_p$. The plumes are dark when seen in projection against the disk and scatter more strongly at shorter wavelengths when seen off the disk[2]. Accordingly we assume the particles to be small, so their cross section for extinction, σ_p, can be written as

$$\sigma_p = \frac{2\pi^2 K}{\lambda} a^3 \qquad (1)$$

where a is the root mean cube of the particle radii, λ the wavelength and K a constant of order unity[9]. Thus, we are considering extinction due to particles with $a < \lambda$, but are ignoring contributions from large particles and Rayleigh scattering $a \ll \lambda$. For an idealised cylindrical volume circumscribing the plume, we then take the optical depth, τ, to be

$$\tau = \sqrt{2}r_p\sigma_p n_p = \frac{2\sqrt{2}\pi^2 K}{\lambda} a^3 r_p n_p \qquad (2)$$

where n_p is the number density of particles.

If we take f_p to be the volume fraction of the cylindrical volume actually filled by plume structure, the total number of particles in the plume, N_p, is

$$N_p = f_p \pi r_p^2 n_p h_p \qquad (3)$$

and the total mass of particles, M_p,

$$M_p = m_p N_p = \frac{4}{3}\pi a^3 \rho_p N_p \qquad (4)$$

where ρ_p is the individual density of the particles.

Solving equation (2) for n_p and substituting in equations (3) and (4) we find that:

$$M_p = \frac{\sqrt{2} f_p \tau}{3K} \rho_p \lambda r_p h_p \qquad (5)$$

which is independent of particle size a.

The resurfacing rate due to n plumes, R (cm s^{-1}), can then be written in terms of M_p and the time of flight of the particles, t_p. If $R \equiv \sum_{p=1}^{n} R_p$, then:

$$R = \frac{\sum_{p=1}^{n} (M_p/t_p)}{4\pi R^2 \rho_p} = \frac{\tau}{6\pi K}\frac{g^{1/2}\lambda}{D^2}\sum_{p=1}^{n} f_p h_p^{1/2} r_p \qquad (6)$$

where D is the diameter of Io and we have used the ballistic flight time given by $t_p \equiv 2\sqrt{2}h_p^{1/2}g^{-1/2}$, g being the acceleration of gravity on Io ($\sim$180 cm s^{-2}). Gas density in the densest parts of the plumes might be enough for drag forces to be important but in the thin parts near the top from which the particles fall back to the surface the ballistic, free fall velocities are supersonic with respect to Io's tenuous atmosphere and the ballistic flight time should be close to the actual flight times of the particles. For example, the ballistic time is about 10^3 s from 100 km altitude.

Taking $\tau \geq 1$, $\rho_p \geq 2$ g cm^{-3}, $K \sim 1$, and $\lambda = 4.8 \times 10^{-5}$ cm, Table 1 gives lower limits for the quantity M_p/f_p for each plume, the total for all plumes and R_p/f_p. Using the same quantities in equation (6) yields a value for R/f, if we assume $f_p \equiv f$ for all plumes.

$$\frac{R}{f} \geq 1.1 \times 10^{-10} \quad \text{cm s}^{-1}$$

The value for R depends on the estimate of f, which must await closer geometric analysis of the images. Assuming that 10% or more of the volume is filled by plume structure we obtain

$$R \geq \sim 3.5 \times 10^{-4} \text{ cm yr}^{-1}$$

This lower limit on the rate is a factor of 100 lower than the lower limit derived from the cratering argument earlier. Our assumptions have been extremely conservative, however. In particular, the derived plume resurfacing rates refer only to a limited size range of particles in plumes actually observed by Voyager 1; larger particles (up to millimetre sizes) could be a significant portion of the mass delivered to the surface. We have also neglected resurfacing due to condensation of the gaseous component of the eruptions, which could at least be equivalent to the particulate mass[10]. Also, considerable resurfacing must be done by volcanic flow activity, for which there is abundant evidence in the images[2,6]. Finally, there are many suggestions in the imaging data of numerous smaller gas/particulate venting below the level detectable as plumes. This low level activity, if constant, could be one of the most important contributors to resurfacing. Taking these factors into account, we conclude that the resurfacing rates required by the crater arguments are in general agreement with the observed level of volcanic activity, but that resurfacing mechanisms in addition to particulate

($a < \lambda$) deposition by the eight known volcanoes are probably necessary unless recent cratering rates at Io have been 100–1,000 times lower than current inner Solar System rates (which we consider unlikely). We will use 10^{-1} cm yr^{-1} as our nominal rate and 3×10^{-4} cm yr^{-1} as an extreme lower limit.

Implications for volatile history

The degree of resurfacing described above has profound implications for the geochemistry and geophysics of Io. It implies that large portions of the crust or interior of Io have been turned over or recycled many times in geologic time. We will now investigate the implications of this rapid resurfacing for volatile production rates and volatile loss from Io.

Knowledge of the volatile composition around Io comes primarily from ground-based studies of neutral and ionic species around Io, in the magnetosphere inside Io's orbit, and from Voyager 1 remote sensing and *in situ* analyses of the plasma torus. Materials identified in the torus include (refs in parentheses), Na I (11), Na nuclei (12), K I (13), S II (14), S III (15), S IV (15), S nuclei (12), O II (15–17), O III (10), O nuclei (12), H (18) and possibly S_2^+ or SO_2^+ (17) (see ref. 19 for a summary table). There are obviously still large gaps in our knowledge of the material in Io's vicinity and the inner magnetosphere. However, it seems clear that Io is the most probable source of most, if not all, of this material. The bulk of the data indicates that S and O are the most abundant species, Na less abundant but still significant and other species present only in trace amounts (leaving aside the issue of hydrogen). Thus, despite possible fractionation effects and observational selection factors, we should look for these elements in the material currently escaping from Io. We can estimate escape rates for some of these species assuming Io is the source. For Na, a rate of $\sim 10^8$ cm^{-2} s^{-1} is required to produce the observed cloud[20,21]. Hydrogen was estimated at 10^{10} cm^{-2} s^{-1} (ref. 18) from Pioneer 10 data and the lack of Lyα emission in the Io torus during the Voyager 1 encounter[15] suggests this is an upper limit at present. Injection of S and O ions into the magnetosphere supplies energy which, if radiated away by the observed UV emission features requires a supply rate of $\sim 10^{10}$ cm^{-2} s^{-1} for these ions[15].

Sulphur has long been regarded as a likely candidate surface material for Io, primarily to explain the very steep decrease in Io's reflectance at short visible wavelengths[22–24]. The volcanism provides an obvious source for sulphur and sulphur compounds, since these are some of the more abundant materials produced in terrestrial volcanism[25–27]. Sulphur flows are proposed as one source of coloured features on Io's surface[28]. Sulphur compounds, particularly S and SO_2, have been suggested as the driving gas for the explosive volcanisms[2,10] and SO_2 has been identified in absorption in infrared spectra from the IRIS experiment on Voyager 1[29]. What is seen in the torus is thus in general agreement with the characteristics of Io and its volcanism (see also ref 19).

We have no direct evidence concerning the water content in Io's volcanism. Water is, of course, a common constituent in terrestrial volcanism (mostly recycled groundwater), and Io may well have had a significant water content at formation[30,31]. However, based on the absence of any indication of the strong IR H_2O absorptions near 3 μm, Pollack *et al.*[32] place a limit of 10% fractional coverage of water ice on Io and 10^{-3} relative abundance of bound water. Since the areas covered by individual plume ejecta range up to ~10% of the observed disk of Io, water condensation over the area affected by the volcanoes should have been observed; it seems that water must be a minor constituent of Io's volcanism or absent altogether.

We can now take the estimated resurfacing rates and the observed supply rates to the torus to draw some conclusions about the history of volatile supply on Io. Assuming an average of 20 AMU for the species involved, resurfacing rates of 3×10^{-4} cm yr^{-1} and 10^{-1} cm yr^{-1} supply $\sim 10^{12}$ atoms cm^{-2} s^{-1} and $\sim 3 \times 10^{14}$ atoms cm^{-2} s^{-1} respectively. It is clear that the current escape rate of $\sim 10^{10}$ cm^{-2} s^{-1} can remove only a fraction of the material being brought to the surface by volcanism.

Table 1 Plume dimensions and resurfacing parameters

Plume no.	r_p* (km)	h_p* (km)	M_p/f_p ($\times 10^9$ g)	R_p/f_p ($\times 10^{-12}$ cm s^{-1})
1	500	280	≥63	≥66.1
2	105	100	≥4.7	≥8.3
3	125	70	≥4.0	≥8.27
4	37	95	≥1.6	≥2.85
5	100	80	≥3.6	≥7.07
6	125	80	≥4.5	≥8.84
7	90	120	≥4.9	≥7.79
8	70	70	≥2.2	>4.63
Total			≥89	113

* From ref. 6.

Io has only a very tenuous atmosphere. Limits of surface pressure range from 10^{-6} bar (stellar occulation) to 10^{-8} to 10^{-11} bar from ionospheric models[33,35]. If only 1% of the material in Io's volcanoes were volatile, then gas would be being supplied at rates of 1 to >200 times greater than the escape rate. Thus, to avoid building up an atmosphere beyond the observed limits, we must conclude that either there are few or no volatiles associated with Io's volcanoes or that virtually all of the volatiles produced are being recycled back into the surface. An extremely low volatile content is unlikely since some gas is needed to drive the observed eruptions[2]. Thus we conclude that whatever gases are being produced are probably being condensed or otherwise reincorporated onto the surface and only minor amounts are escaping from the volcanoes directly or from a tenuous atmosphere.

The observation of SO_2 gas over a volcanic plume[29] and the possibility of an atmosphere containing SO_2 in equilibrium with solid SO_2 at some temperature[19,29] are consistent with the above arguments. In addition condensed volcanic gases explain some of the puzzling features of Io's spectrum[32,36]. Hapke[37] has suggested, based on transmission spectra of various volcanic materials, that H_2S, SO_2 and possibly S_2O might explain some of the features in Io's reflection spectrum. Fanale *et al.*[38] and Smythe *et al.*[39] have demonstrated that diffuse reflection spectra of SO_2 frost explain the absorption near 4.0 μm very well.

Some of the albedo features on Io's surface may also be understood in terms of condensation phenomena. Inspection of Fig. 12, 14*a*, 16, 17*a* and 18 in ref. 2 shows that the deposits of small particles or ash is dark and highly localised. We may match this process with the more localised bright rings outside the inner cores of volcanically active areas (Fig. 18*a* and *b* in ref 2). We propose that the bright rings are formed on the outer edges of hot volcanic areas where the surface temperature is low enough to allow sulphur and/or SO_2 frost to deposit from the volcanic gas with the outer limit being set by the outer limit of higher vapour density associated with the plume. An interesting example is the dark area south of Plume 2 shown in *Plate 1*, p. 782 see Strom *et al.*[7]. The bright deposit from this plume does not cross this dark 'lava lake.' We suggest that this is due to high temperature of this feature[40] preventing condensation of SO_2.

An important aspect of Io's volcanism is that is has probably been going on for $(4–4.5) \times 10^9$ yr (ref. 41). The above range of inferred resurfacing rates implies the supply of anywhere from ~0.1 to ~10 times the total mass of Io over geological time. This is another way of viewing the conclusion reached earlier that significant amounts of Io's crust and interior have been recycled several times. The current escape rates, on the other hand, only amount to a loss of 10^{-4} of the mass of Io over geologic time. Thus the currently observed rates do not imply significant loss of material from Io since formation. The presence of abundant heavier volatiles such as sulphur is not therefore particularly surprising since only 1% of Io's total sulphur content (if near solar abundance) would have been lost at these rates. However, the ease with which escape occurs may be a strong function of the volatiles involved, and escape rates could have been very different earlier in Io's history. In fact it seems likely that the large volumes of material brought up by the volcanic activity must have originally contained larger amounts of volatiles such as H_2S and H_2O which were lost through dissociation, ionisation and H escape. This would explain the apparent absence of H_2O combined with active current degassing of sulphur compounds (see also ref. 19).

Implications for supply to plasma and neutral torus

The observation of volcanic eruption plumes on Io also raises the question of what processes are responsible for injecting material into the neutral clouds and plasma torus associated with Io. An initial question is whether material is directly escaping in the eruption process. The height of the observed plumes suggests ejection velocities of $\sim 1\ \mathrm{km\ s^{-1}}$, below the $2.5\ \mathrm{km\ s^{-1}}$ escape velocity. Since particulate material in an eruption is entrained with the gas, particle velocities greater than the gas velocity are unlikely. Velocities of $1\ \mathrm{km\ s^{-1}}$ are already high for this type of eruptive activity, and higher velocities are unlikely[10]. Thus it seems that direct ejection of particulate material with velocities greater than the escape velocity is not responsible for supplying material to the magnetosphere.

Atmosphere loss by either thermal or non-thermal mechanisms of gas supplied by the volcanoes is potentially a major source for the observed escape flux. Jeans escape of sulphur from an atmosphere with a surface pressure of 10^{-10} bar, for instance, would supply $\sim 10^{10}$ atoms $\mathrm{cm^{-2}\ s^{-1}}$ from the critical level of ~ 500 km if the exospheric temperature were 1,200 K and sulphur were a major constituent at that level. Note that the volcanic eruptions may supply material to altitudes near to or even above the exobase. Thus the volcanic activity may bring gases to a level where they will be exposed to escape by thermal or non-thermal mechanisms.

Non-thermal escape mechanisms such as magnetospheric sweeping of ions or dissociative recombination reactions have been suggested as strong possibilities in Io's environment[42–44]. These may dominate thermal escape, and here again the volcanic eruptions may play a major role. If Io is 'protected' by the interaction of the co-rotating plasma with a possible Io magnetic field[45], then the volcanoes may on some sides of Io supply material above the stand-off altitude, allowing interaction with and sweeping by the magnetosphere. Also, although direct escape of particulates from the volcanoes is unlikely, it is possible that some of the very fine micrometre or submicrometre particles in the volcanic plumes may become charged in the plasma environment and be removed by magnetic sweeping. This is at least potentially an important source of dust in the inner magnetosphere. Direct sputtering (see ref. 42) of dust in the plumes by magnetospheric particles is also possible, although the surface area exposed in plumes is less than Io's surface area.

The processes described above may be sufficient to account for current losses of sulphur and oxygen to the torus and for the escape of hydrogen earlier in Io's history. The problem of sodium escape is more difficult, at least partially because we have considerably more information about the neutral species, whereas we only observe the magnetospheric ions for most other species. Three points particularly deserve comment: the apparent constancy of sodium supply since 1973, the observed velocity distribution of the sodium atoms, and the asymmetry of escape required by the cloud geometry.

A constant supply of sodium by volcanoes is not ruled out by the observed volcanic activity, which certainly suggests a rather high average level of activity. However, considering the short lifetime against ionisation implied by observations (20–50 h, ref. 46), major fluctuations in volcanic activity (by a factor of ≥ 2) should have been observed in the sodium data if the sodium escape is directly linked to the volcanoes.

Hapke[37] has suggested that Na (and K) are supplied to the torus by evaporation from hot silicate lavas ($T \sim 1{,}450\ K$), possibly carried to great altitudes in volcanic plumes. This seems unlikely for several reasons. First, no temperature of this magnitude has been observed on Io[40], and models for explosive gas volcanism start with much more modest temperatures[10]. Second, any hot silicate which could be ejected by such a gas eruption would cool considerably during its 10 min trip to the top of a plume. Finally, the large Doppler width of the sodium emission lines ($\sim 5\ \mathrm{km\ s^{-1}}$ FWHM) and the high velocity asymmetry discovered by Trafton are evidence against thermal escape[47,48], even if the volcanic activity supplies sodium or sodium compounds to the atmosphere. Since only neutral sodium is involved, non-thermal mechanisms involving ionisation are not possibilities.

Sputtering of atoms from the surface by impacting magnetospheric ions has been suggested as the sodium supply mechanism by Matson and coworkers[42]. Before Voyager, the primary difficulties with this mechanism were the very low efficiencies for sputtering by protons (although ice may be an exception[49]) which were then thought to be the major positive ions in the magnetosphere, and the possible thermalisation of sputtered atoms in Io's atmosphere, requiring yet another energetic event to supply escape energy. The discovery of large densities of sulphur and oxygen ions in the inner magnetosphere has removed the difficulty with yields. For a density of $10^3\ \mathrm{cm^{-3}}$ the flux of ions at Io's position in the co-rotational direction is $\sim 6 \times 10^9\ \mathrm{cm^{-2}\ s^{-1}}$. Whether these ions impact Io directly or are precipitated into it by an Io magnetosphere, the energy of the impact will be comparable to the co-rotational energy, ~ 500 eV for sulphur at Io's orbit. Heavy ions in this energy range have substantial sputtering yields of the order of unity or greater[50]. Ions of this energy can also withstand several collisions with atmospheric molecules without losing a significant fraction of their energy. Thus the known magnetospheric ion population is sufficient to supply the required sodium flux even for modest sodium abundances in the surface (1–10%).

The major issue remaining is whether sputtered atoms escape Io or are merely supplied to a thermalising atmosphere for later escape. An SO_2 atmosphere (or any other atmosphere that meets the criteria discussed above for volcanic gases that can be reincorporated in the surface) may be considerable at night[19,29]. Thus this class of atmospheric model may allow atmospheric escape of gases in the daytime and sputtering of surface material at night. Note that sulphur should also be sputtered in some quantity (along with other surface materials), so that volcanic supply of daytime gases might represent a modulation on a lower level, a more steady supply rate.

The shape of the cloud and the velocity profile asymmetry seem to require asymmetric escape of neutrals from Io, most likely from the inner hemisphere, either the trailing or the leading portion thereof[51,52]. This asymmetry could in principle arise either from asymmetric source or sink processes. The volcanoes are more or less uniformly distributed in longitude[2,6], and the night/day effect suggested above averages out to uniform sputtering over an orbit if no other processes operate. Interaction with the magnetosphere remains the best candidate for producing this asymmetry either by creating preferential regions for ion precipitation or possibly by shielding neutrals from electron ionisation in regions inside an Io 'magnetosphere'[45]. It is possible that a day/night modulation combined with a magnetospheric asymmetry could produce the east–west sodium anomalies reported by Bergstralh *et al.*[53] and Goldberg *et al.*[54], but more detailed models will be necessary to study these possibilities.

Conclusion

The resurfacing rates implied by observations of active volcanism on Io are sufficient to explain the lack of recognisable impact craters on Io's surface. The implied rates of material supply to Io's surface require major recycling of Io's upper crust and mantle and that volcanic volatiles be condensable under Io's surface conditions, consistent with the observation of SO_2[29] and models of Io's volcanism[10]. Atmospheric escape of a small fraction of the volcanic gas can supply the S and O in Io's torus and these same species, as energetic heavy ions, may be responsible for the supply by sputtering of high velocity neutral sodium.

We thank our colleagues on the Voyager imaging science team, S. Kumar and J. Pearl of the Voyager IRIS team and R. W. Carlson and D. L. Matson of JPL for stimulating discussions. We also thank F. P. Fanale (JPL) for helpful criticisms. This article presents results of one phase of research carried out at JPL under NASA contract NAS7-100.

Received 16 July; accepted 3 August 1979.

1. Morabito, L. A. *et al. Science* **204**, 972 (1979).
2. Smith, B. A. *et al. Science* **204**, 951–972 (1979).
3. Shoemaker, E. M. in *Impact and Explosion Cratering*, (eds Roddy, D. J., Pepin, R. O. & Merrill, R. B.) 617 (Pergamon, New York, 1977).
4. Wetherill, G. W. in *Impact and Explosion Cratering*, (eds Roddy, D. J., Pepin, R. O. & Merrill, R. B.) 613 (Pergamon, New York, 1977).
5. Soderblom, L. A. in *Impact and Explosion Cratering* (eds Roddy, D. J., Pepin, R. O. & Merrill, R. B.) 629 (Pergamon, New York, 1977).
6. Carr, M. H. *et al. Nature* **280**, 729–733 (1979).
7. Strom, R. G., Terrile, R., Masursky, H. & Hansen, C. *Nature* **280**, 733–736 (1979).
8. McCauley, J. F. & Soderblom, L. A. *Nature* **280**, 736–738 (1979).
9. van de Hulst, H. C. *Light Scattering by Small Particles*, 179–183 (Wiley, New York, 1957).
10. Smith, B. A., Shoemaker, E. M. & Cook, A. *Nature* **280**, 738–743 (1979).
11. Brown, R. A. in *Exploration of the Planetary System* (eds Woszczyk, A. & Iwaniszewski, C.) 527 (Reidel, Dordrecht, 1974).
12. Vogt, R. E. *et al. Science* **204**, 1003 (1979).
13. Trafton, L. *Nature* **258**, 790 (1975).
14. Kupo, I., Meckler, Y. & Eviatar, A. *Astrophys. J. Lett.* **205**, L51 (1976).
15. Broadfoot, A. L. *et al. Science* **204**, 979–982 (1979).
16. Pilcher, C. B. & Morgan, J. S. *Science* (in the press).
17. Bridge, H. S. *et al. Science* **204**, 987–991 (1979).
18. Carlson, R. W. & Judge, D. L. *Icarus* **24**, 395 (1975).
19. Kumar, S. *Nature* **280**, 758–760 (1979).
20. Brown, R. A. & Yung, Y. L. in *Jupiter*, (ed. Gehrels, T.) 1102 (University of Arizona Press, 1976).
21. Smyth, W. H. & McElroy, M. B. *Planet. Space Sci.* **25**, 415 (1977).
22. Wamsteker, W., Kroes, R. L. & Fountain, J. A. *Icarus* **23**, 417 (1974).
23. Fanale, F. P., Johnson, T. V. & Matson, D. L. *Science* **186**, 922 (1974).
24. Nash, D. B. & Fanale, F. P. *Icarus* **31**, 40 (1977).
25. Bunsen, M. R. *Ann. Chim. Phys.* 3rd Ser. **38**, 215 (1853).
26. Lacroix, A. *Bull. Soc. Min. Fr.* **30**, 219 (1907).
27. Stoiber, R. E. & Rose, W. I. Jr, *Geochim. cosmochim. Acta* **38**, 495 (1974).
28. Sagan, C. *Nature* **280**, 750–753 (1979).
29. Pearl, J. *et al. Nature* **280**, 755–758 (1979).
30. Fanale, F. P., Johnson, T. V. & Matson, D. L. in *Planetary Satellites* (ed. Burns, J. A.) 379 (University of Arizona Press, 1977).
31. Pollack, J. B. & Reynolds, R. T. *Icarus*, **21**, 248 (1974).
32. Pollack, J. B. *et al. Icarus* **36**, 271 (1978).
33. Smith, B. A. & Smith, S. A. *Icarus*, **17**, 218 (1972).
34. Brown, R. A. & Yung, Y. L. in *Jupiter* (ed. Gehrels, T.) 1102 (University of Arizona Press, 1976).
35. Johnson, T. V., Matson, D. L. & Carlson, R. W. *Geophys. Res. Lett.* **3**, 293 (1976).
36. Cruikshank, D. P., Jones, T. J. & Pilcher, C. B. *Astrophys. J. Lett.* **225**, L89 (1978).
37. Hapke, B. *Geophys. Res. Lett.* (submitted).
38. Fanale, F. P. *et al. Nature* **280**, 761–763 (1979).
39. Smythe, W., Nelson, R. & Nash, D. B. *Nature* **280**, 766 (1979).
40. Hanel, R. *et al. Science* **204**, 972–976 (1979).
41. Peale, S. J., Cassen, P. & Reynolds, R. T. *Science* **203**, 892 (1979).
42. Matson, D. L., Johnson, T. V. & Fanale, F. P. *Astrophys. J. Lett.* **192**, L43 (1974).
43. Johnson, T. V., Matson, D. L. & Carlson, R. W. *Geophys. Res. Lett.* **3**, 293 (1976).
44. Cloutier, P. *et al. Astrophys. Space Sci.* **55**, 93 (1978).
45. Kivelson, M. G., Slavin, J. A. & Southwood, D. J. *Science* (in the press).
46. Carlson, R. W., Matson, D. L. & Johnson, T. V. *Geophys Res. Lett.* **2**, 469 (1975).
47. Trafton, L. *Astrophys. J. Lett* **202**, L107 (1975).
48. Carlson, R. W. *et al. Astrophys. J.* **223**, 1082 (1978).
49. Brown, W. L. *et al. Phys. Res. Lett.* **40**, 1027 (1978).
50. Carter, G. & Colligon, J. S. *Ion Bombardment of Solids*, Ch. 7, 310–353 (Elsevier, New York, 1968).
51. Carlson, R. W. *et al. Astrophys. J.* **223**, 1082 (1978).
52. Matson, D. L. *et al. Science.* **199**, 531–533 (1978).
53. Bergstralh, J. T., Matson, D. L. & Johnson, T. V. *Astrophys. J. Lett.* **195**, L131 (1975).
54. Goldberg, B. A. *et al. Bull. Am. astr. Soc.* **10**, 579 (1978); *EOS* **59**, 1122 (1978).

Sulphur flows on Io

Carl Sagan

Laboratory for Planetary Studies, Cornell University, Ithaca, New York 14853

The colour patterns observed on Io indicate extensive surface flows of quenched molten sulphur

Io has the reddest surface in the Solar System, with the possible exception of Amalthea, which may in fact, be coloured by material lost from Io. Io also displays the largest variation of colour with planetocentric longitude of any other object in the Solar System[1]. Voyager 1 close-encounter photomosaics have clarified both the coloration of Io and its surface variation (see, for example the colour images in Smith *et al.*[2]). Although the Voyager 1 photometric fidelity is not better than 10%, general colour patterns are clear. The surface is mottled with dozens of large, and, probably, thousands of small randomly distributed dark patches, which in most cases are adjacent to red or orange coloured terrain. Equatorial and temperate latitudes are rich in these bright colours. Polar terrain tends to be darker, confirming earlier groundbased impressions. There are also extensive yellow, and white or blue-white, provinces on the satellite's surface. Many of the dark spots are active or recently active shallow volcanic constructs[2–6]; seven of the volcanoes have been observed by Voyager 1 in violent eruption. The pattern of albedo features points strongly to a succession of complex effusive and fluvial events. Some dark localised diffuse features exhibit radiating patterns implying downslope flow, such as in *Plate* 2, p 783. In this image there is a hint of sinuous anastamosing channels, ~300 km long, which are blood-red with some traces of black material down the principal channel. One channel is red proximate to the caldera, fading to yellow distally. In general, dark flows are short and diffuse, red flows longer, and orange and yellow flows tend to be distributed in sheets of considerable area.

A crude similarity between the integrated-disk colour of Io and the spectrum of elemental sulphur led Wamstecker *et al.*[7] to suggest the presence of abundant sulphur on the ionian surface. Sulphur has also been identified in the Io-related circumjovian torus both by Earth-based observations[8] and by Voyager 1 experiments[9,10]; SO_2 has been identified in the gas phase by Voyager 1 IR spectroscopy[11]; and it has been suggested that SO_2 frost may explain some of the broadband IR spectral features in the integrated disk of Io[12,13]. After water and CO_2, sulphur and its compounds are generally the first or second most abundant effluents in terrestrial volcanoes, and fumaroles, in some cases comprising tens of per cent by mass[14]. Where fumarole temperatures are $\lesssim 650$ °C, sulphur gases tend to be the major constituents after steam, and the corresponding volcanic constructs are called solfataras. Sulphur and its compounds have been thought[2,15] to drive the volcanic plumes because of the absence on Io, unlike the other galilean satellites, of surface water frost in groundbased IR spectroscopic observations. Io has been substantially devolatilised. H_2S is rapidly photodissociated and, because of its very low freezing point, hydrogen has an opportunity to escape efficiently from Io by processes such as Jeans evaporation, or magnetospheric sweeping. As Io becomes more and more oxidising, S_n and SO_2 become prime candidates for the principal volcanic effluents. Ionian effusive flows and sulphur volcanoes may be driven by the lithostatic pressure gradient and explosive decompression of volatiles[16], with the energy derived from tidal dissipation dependent on the resonance lock with Europa[17]. However, the following discussion is independent of the mechanics of Io volcanism.

Sulphur coloration

There are perhaps dozens of sulphur allotropes with differing physical, chemical and optical properties derived from differences in both the packing and the bonding of polyatomic sulphur; their melting points vary from 110 to 119 °C. Because the melting point of sulphur is ~1,000 °C less than that of

0028-0836/79/350750—04$01.00