# THERMAL STUDIES OF PLANETARY SURFACES

David Morrison
University of Hawaii
Honolulu, Hawaii 96822

## ABSTRACT

Ground-based and spacecraft observations of planets, satellites, and asteroids in the thermal infrared have provided a wealth of information on planetary temperatures, dimensions, and surface properties. Internal heat sources have been revealed for Jupiter, Saturn, and Neptune, and active volcanism on Io has been discovered and monitored. The thermophysical properties of Mars have been mapped for nearly all the surface by spacecraft, and ground-based observations have given similar information for the Galilean satellites of Jupiter. Infrared radiometry thus sheds important light on significant problems of dynamics, interiors, and surfaces of solar system bodies.

## I. INTRODUCTION

The brightest infrared sources in the sky are in the solar system, and studies of this radiation have played an important role in infrared astronomy since the pioneering thermocouple measurements of the 1920s by Coblentz and Lampland at Lowell and Pettit and Nicholson at Mt. Wilson. Indeed, the history can be traced back even further, to the detection of lunar thermal radiation in 1869 by the Earl of Rosse with his 3-foot reflector. The temperature measurements from the premodern era are reviewed by Pettit (1961) and Sinton (1961); the primary results of interest, beyond the simple measurement of planetary temperatures, were the discovery that there was no diurnal infrared temperature variation on Venus, the determination from eclipse cooling and heating curves that the lunar surface has an exceedingly low thermal inertia, and the measurements of the diurnal temperature variations on Mars.

The rapid improvements of the 1960s in infrared instrumentation led to the birth of non-solar-system infrared astronomy. Initially, the bright planetary sources served as calibration standards for other objects, and they are still used for this purpose at the longer



*C. G. Wynn-Williams and D. P. Cruikshank (eds.), Infrared Astronomy, 89–106.*

wavelengths. However, no truly satisfactory standard has been found, since the planets are generally variable, often have pronounced spectral structure, and have such large angular sizes that they overfill the beam of most photometers. Mars has been used frequently, but it is perhaps the least appropriate calibrator, since its thermal emission can vary substantially with weather as well as in ways more readily modeled. Potentially the most useful solar system standards are the larger asteroids, such as Ceres, which should behave in a predictable way and are easily measured by existing systems at wavelengths from 8 μm to beyond 100 μm. However, today the calibration situation is usually reversed, with hot stars providing the primary reference, and the planets being studied for their own sakes.

Ground-based and spacecraft thermal observations, primarily broad-band photometry, continue to occupy a central place in planetary investigations. Among the highlights of the past decade have been the measurements of the internal energy sources of Jupiter, Io, Saturn, and Neptune, the discovery of temperature inversions in the upper atmospheres of the Jovian planets and the analysis of hydrocarbon trace chemistry; the use of thermal spectra to sound the temperature-pressure structure of planetary atmospheres; measurement of the composition of planetary atmospheres, including the fundamental datum of the hydrogen-to-helium abundance ratio; the identification of the composition of the Martian polar caps; the measurement of thermal inertia for numerous planets and satellites and the systematic mapping of inertia over the surfaces of the Moon and Mars; and the measurement of the sizes and albedos of more than a hundred asteroids and nearly a dozen satellites too small to be resolved. In this chapter I discuss four topics: the energy balance of the Jovian planets; the thermal inertias of surfaces; asteroid diameters and albedos, and the volcanic activity of Io. The composition and structure of planetary atmospheres are discussed elsewhere in this volume by Orton and by Encrenaz and Combes.

## II. INTERNAL HEAT SOURCES OF THE JOVIAN PLANETS

One of the most exciting and unexpected early discoveries of the modern era of infrared astronomy was the existence of large internal heat sources in Jupiter and Saturn. A Jovian excess was first suggested by Low (1966), based on the high brightness temperatures measured at 10 and 20 μm, and this excess was confirmed for Jupiter and extended to Saturn through Lear Jet observations extending out to nearly 100 μm (Aumann et al. 1969). These measurements demonstrated that Jupiter radiated more energy than it received from the Sun, but they could not establish the magnitude of the internal heat source since it was not possible to measure the radiation from the night side or from the polar regions. Spacecraft observations from a variety of directions are required to establish a model-independent value for the heat source. In the case of Saturn, the situation has been further complicated by the presence of thermal emission from the rings.

https://doi.org/10.1017/S0074180900075148 Published online by Cambridge University Press

The first spacecraft measurements of thermal radiation from Jupiter were made from Pioneer 10 in 1973 and Pioneer 11 in 1974. Pioneer 10 made the first observations of the night side of the planet, and in addition Pioneer 11 was able to contribute data on the polar temperatures. The descriptions of the individual encounters are given by Chase et al. (1974) and Ingersoll et al. (1975a), with a summary of the results of both missions by Ingersoll et al. (1975b). Although temperature differences were measured between bands and zones, there was no significant cooling toward the poles or at high phase angles, leading to a derived global effective temperature of 125 $\pm$ 3 K, marginally lower than the Earth-based values. Independently calibrated infrared observations from the Voyager Jupiter encounters in 1979 are consistent with the Pioneer results. The source of the internal energy of Jupiter is apparently primarily primordial, representing the slow leakage of heat from a still hot interior. Models of the early evolution of the planet by Bodenheimer (1974), Graboske et al. (1975), and others indicate that the present heat source of about 4 x $10^{17}$ watts is easily fitted by standard evolutionary models for the planet.

In the case of Saturn, all of the early infrared observations were made at a time when the rings were widely open and emitting with a brightness temperature of 90-95 K (Morrison 1974a). In large aperture photometry, the rings contributed as much radiation as the planet itself. During the late 1970s the rings closed, lowering both their temperature and their solid angle, while infrared models of the rings (e.g., Cuzzi 1978) permitted corrections to be made for their contribution. Thus Stier et al. (1978) derived an effective temperature for the planet of about 90 K, Erickson et al. (1978) observed 97 K, and Courtin et al. (1979) obtained 95 K, all suggestive of a heat source somewhat smaller than that of Jupiter. In 1979, the Pioneer 11 flyby of Saturn provided an opportunity for more comprehensive measurements, made with the same radiometer used five years earlier to observe Jupiter. The effective temperature found by the Pioneer investigators was 97 $\pm$ 3 K (Ingersoll et al. 1980, as corrected by G. Orton, private communication). The implied internal power of 2 x $10^{17}$ watts is substantially larger than expected from evolutionary modeling of the sort that worked for Jupiter (Pollack et al. 1977; Pollack 1978). Apparently an energy source is needed in addition to primordial heating, with the most frequently suggested candidate gravitational separation of hydrogen and helium in the core.

Jupiter and Saturn each radiate most of their thermal energy in spectral regions accessible from the ground, but for Uranus and Neptune the peak of the spectrum moves beyond 30 μm, so that ground-based photometry is less readily interpreted. The first evidence for an internal heat source was presented by Morrison and Cruikshank (1973a) and Rieke and Low (1974), who found Neptune to be brighter than Uranus in the 20-30 μm band in spite of its greater distance from the Sun. This result was much strengthened by observations extending to longer wavelengths obtained from the Kuiper Airborne Observatory

(KAO) by Loewenstein et al. (1977a,b) and from balloon altitudes by Stier et al. (1978). The work clearly showed that Uranus has very nearly the temperature expected for a rapidly rotating planet in equilibrium with the insolation (58 K), while Neptune has essentially the same effective temperature (56 K), and is apparently emitting about twice as much energy as it absorbs from the Sun. The reason for the difference between the two planets is unknown, although it has been suggested by Trafton (1974) that tidal effects from Triton may heat the interior of Neptune.

The current information on heat sources in the giant planets is summarized in Table 1. Although Jupiter has the highest total luminosity, it is interesting that the luminosity-to-mass ratio is greater for Saturn, presumably as a result of an additional energy-producing mechanism in its core. The heat source in Neptune is large only in comparison to its feeble illumination by the Sun; in absolute units it is two orders of magnitude below Jupiter and Saturn, and in specific units one order of magnitude lower. The observational upper limit for an intrinsic luminosity for Uranus is a factor of 2 or 3 below the value derived for Neptune.

TABLE 1

INTERNAL HEAT SOURCES OF THE JOVIAN PLANETS

| | | |
|---|---|---|
| JUPITER | $T_B$ | $= 127 \pm 3$ K |
| | Total power | $= (2.0 \pm 0.2)$ solar input |
| | Internal power | $= 7$ W m$^{-2}$ $= 4 \times 10^{17}$ W |
| | | $= 0.2$ μW/ton |
| SATURN | $T_B$ | $= 97 \pm 3$ K |
| | Total power | $= (2.8 \pm 0.4)$ solar input |
| | Internal power | $= 3.5$ W m$^{-2}$ $= 2 \times 10^{17}$ W |
| | | $= 0.3$ μW/ton |
| URANUS | $T_B$ | $= 58 \pm 3$ K |
| | Total power | $\simeq$ solar input |
| | Internal power | $< 0.1$ W m$^{-2}$ $= 10^{15}$ W |
| | | $< 0.01$ μW/ton |
| NEPTUNE | $T_B$ | $= 56 \pm 3$ K |
| | Total power | $= (2.5 \pm 0.5)$ solar input |
| | Internal power | $= 0.4$ W m$^{-2}$ $= 3 \times 10^{15}$ W |
| | | $= 0.03$ μW/ton |

https://doi.org/10.1017/S0074180900075148 Published online by Cambridge University Press

All of these measurements of the infrared luminosity of the Jovian planets provide important constraints on their internal structure and thermal evolution. Only for Jupiter does the mechanism that produces the heat appear to be well understood, and even there the processes of energy transfer that produce polar regions as warm as the solar-heated equator are obscure. A great deal of theory remains to be worked out before we can claim to understand these observations.

## III. THERMOPHYSICS OF PLANETARY SURFACES

One of the earliest results of infrared astronomy was the determination of the low thermal conductivity of the upper few centimeters of the lunar crust. As is well known, the response of the surface temperature of a semi-infinite, homogeneous slab of material with temperature-independent thermal properties to changing radiative boundary conditions at its upper surface is a function only of the composite parameter $(K\rho c)^{1/2}$, where K is the thermal conductivity, $\rho$ is the density, and c is the heat capacity. This parameter is called the "thermal inertia", by analogy with mechanical inertia. The greater the thermal conductivity, the more slowly the surface temperature can respond to a rapidly changing insolation, and thus the more "inertia" the surface has. Small inertia, then, implies low conductivity and rapidly fluctuating temperatures.

This paper follows the general practice of using I to represent thermal inertia and expressing it in units of $10^{-3}$ cal cm$^{-2}$ s$^{-1/2}$ K$^{-1}$, which is equal to 41.84 J m$^{-2}$ s$^{-1/2}$ K$^{-1}$. In these units, the typical thermal inertia of the Moon is $I \simeq 1$ and of a solid terrestrial rock is $I \simeq 40$.

To determine thermal inertia, one must measure the response of surface temperature to changing insolation. Nature frequently provides two time scales for these changes, one corresponding to the diurnal cycle and one to brief interruptions of sunlight during eclipses. With each time scale is associated a characteristic skin depth related to the thickness of the surface layer that experiences these transient temperature changes. Typically for low-conductivity surfaces, these depths are from a few millimeters to several centimeters. Measurements of thermal inertia thus characterize the physical properties of the uppermost layer of the regolith and are sensitive to conditions on a scale substantially smaller than can be imaged remotely.

The first application of thermal inertia measurements outside the Earth-Moon system was to Mars, using diurnal temperature variations observed by Sinton and Strong (1960). Morrison (1968) used these data to estimate I = 4-6, with suggestion of a larger inertia for dark areas. Most determinations of the thermal inertia, however, have been made from eclipse cooling and heating observations, generally of unresolved objects.

The Galilean satellites of Jupiter undergo eclipses often and are fairly easily observed with Earth-based telescopes. At the very start of the modern era of infrared astronomy, Murray et al. (1965) observed part of the eclipse cooling curve of Ganymede and concluded from the rapid changes in temperature that its thermal inertia must be lunar-like. Morrison et al. (1971) followed the temperature change for Ganymede over its full range and concluded that I was definitely lower than for the Moon. A series of eclipse measurements made at the Hale 5-meter and at Mauna Kea were summarized by Hansen (1973) and Morrison and Cruikshank (1973b), who found that for all the Galilean satellites $I \simeq 0.3$, indicating a regolith of extremely low conductivity. Radiometric eclipse measurements of Mars' satellite Phobos obtained from Mariner 9 by Gatley et al. (1974) yielded a comparably low thermal inertia.

The advent of spacecraft radiometry in the late 1960s allowed the thermal inertia to be mapped over the surface of a planet with good spatial resolution. The Mariner 6 and 7 flybys of Mars established that the seasonal polar caps were composed of $CO_2$ (Neugebauer et al. 1971). The Mariner 9 orbiter provided coverage of more than 35% of the surface with a resolution of 100 km and established $I = 7$ as an appropriate mean value to characterize the planet, with a range in thermal inertia from 4 to 17 (Kieffer et al. 1973). The much larger value of inertia for Mars as compared with the Moon or the Galilean satellites is a consequence of the atmosphere, which provides additional energy transfer between grains as well as compaction of the surface material by wind.

In 1974, Mariner 10 flew past the night side of Mercury and spatially resolved temperatures were measured. The thermal inertia derived by Chase et al. (1976) from these data ranged between 1.5 and 3.1; the fact that these are higher than the lunar values probably represents in part a radiative contribution to the thermal conductivity at the much higher average subsurface temperatures on Mercury.

The most important application of thermal inertia measurements to date has resulted from several years of systematic orbital observations of Mars carried out with the Viking IRTM (Infrared Thermal Mapper). This multi-channel radiometer was able to map surface areas at a variety of spatial scales and at many points in the diurnal cycle. As the seasons passed on Mars and dust storms developed and subsided, it was also possible to learn how the atmosphere influenced surface temperature. The wealth of information in these observations has emerged in a series of papers beginning with Kieffer et al. (1977), and the analysis is still underway.

In most of the earlier IRTM papers (e.g., Zimbelman and Kieffer 1979), the thermal inertia of surface elements was derived from the predawn temperature; the more the surface cools during the Martian night, the lower the conductivity and the smaller the inertia. (This provides a convenient way to remember the relationship: small

T => small K => small I). The analysis based on predawn temperature alone neglects the effects of albedo on temperature. Later papers (e.g., Palluconi and Kieffer 1981) used the full data set to derive the thermal inertia directly for each spot observed. The results are thereby improved, but not changed in any major way.

Figure 1, taken from Palluconi and Kieffer (1981), is a thermal inertia map of Mars between 60°N and 60°S with a spatial resolution of 2° in latitude and 2° in longitude. The data were sufficient to derive the inertia for 10,171 of the possible 10,800 areas so defined. All of these inertias correspond to clear conditions well separated in time from the major planet-wide dust storms. The total range in derived inertia is $1 \leq I \leq 15$. The distribution of inertias is bimodal, with all values less than 4 associated with northern-hemisphere bright regions (Tharsis-Amazonis, Elysium, and near 330°W, 15°N) comprising 20% of the surface. These regions of low conductivity are probably blanketed by fine deposits of wind-blown dust. The highest inertias generally correspond to dark regions; apparently insulating dust blankets are not composed of dark materials. Probably there are exposures of bare rock in the high-inertia areas, although the observed maximum value of I at this resolution is not high enough to indicate any 2° x 2° area with predominantly bare rock.

One interesting conclusion from the global thermal inertia mapping of Mars relates to the nature of the two Viking lander sites. These sites were selected for smoothness, but both turned out to be rocky: 8% of the surface is rock-covered at VL-1, and 14% at VL-2. The thermal inertias in these areas as shown in Figure 1 are 9 and 8, respectively. This places the landing sites in the higher 20th percentile of the observed inertias. To the extent that thermal inertia measures the extent of bare rocks, it would appear that both sites are atypically rocky. Palluconi and Kieffer (1981) further argue that the two sites have unusually high albedo for their thermal inertias, suggesting they might not be at all like most of the surface. This is a point worth remembering, since it seems likely that the perception of the Martian surface by scientists and lay persons alike will be based for the next generation on pictures taken by these two Viking landers.

The thermal inertia is a quantity that stands midway between a set of temperature measurements and a true characterization of the physical properties of a surface. Even when the observed temperature variations are matched well with a homogeneous heat conduction model, the actual conductivity and porosity of the surface may not be well established. When the data are sparse, we are content to characterize an entire planet by the simple parameter I, but in the case of Mars, it is clear that the surface and the processes modifying it are extremely complex. Thermal inertia mapping is just one of many inputs to developing an understanding of the regolith of a planet.

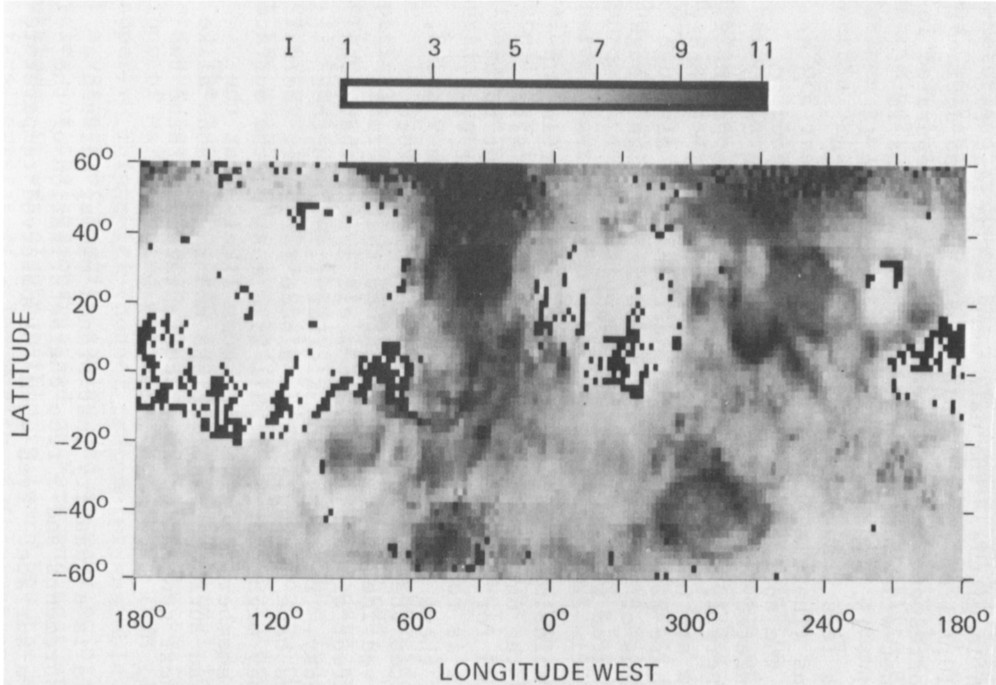


Figure 1: Thermal inertia map of Mars derived from Viking Orbiter 20-μm infrared mapping by Palluconi and Kieffer (1981). The data, all taken under clear atmospheric conditions ($L_s$ 344° to 125°), are analyzed in bins 2° x 2° in size; there are 10,800 such bins in the figure. The bins (5.8% of the total) with no determination of thermal inertia are black. For the other regions, white represents inertia of 1 and dark grey inertia greater than 10 (in units of $10^{-3}$ cal $cm^{-2}$ $s^{-1/2}$ $K^{-1}$), as indicated by the grey scale at the top of the figure. The three large northern-hemisphere areas of low inertia are, from the left, the high albedo regions Tharsis-Amazonis, Elysium, and an unnamed region near longitude 330°. Prominent areas of high inertia include the Viking landing sites in Chryse Planetia (22°.5 N, 47°.8 W) and in Utopia Planetia (48°.0 N, 225°.6 W).

## IV. DIAMETERS AND ALBEDOS OF MINOR PLANETS AND SATELLITES

In solar system applications as elsewhere, infrared photometry is not generally obtained in order to derive a simple temperature. One of the most productive infrared programs of the past few years has involved the measurement of thermal emission from bodies so small that no diameter is known, and thus no brightness temperature can be defined. Instead, the infrared photometry is combined with visible photometry to derive the albedo and diameter of the object.

This radiometric/photometric technique for determining sizes and albedos of small airless objects was first applied by Allen (1970) to measure the diameter of asteroid 4 Vesta, and most of the subsequent applications of this method have been in asteroid studies. The basic principle is simple. For an object of a given angular size in equilibrium with the insolation, the brightness in reflected light will scale with the albedo, while the thermal reradiation of absorbed sunlight scales in the opposite manner, being roughly proportional to (1 - albedo). Measurements of the reflected and the reradiated energies determine the albedo and the size. In practice, the interpretation is model dependent and requires some knowledge of the photometric properties and the infrared emissivity of the object. In addition, there should be a calibration in terms of objects of known size, since the basic uncertainties in the absolute calibration of the magnitude scales also introduce errors. These calibrations, and the details of the thermal models used to interpret the data, are described in detail by Matson (1971), Morrison (1973), Jones and Morrison (1974), Matson et al. (1978), and Morrison and Lebofsky (1979).

Although it might seem in general that the albedos and diameters derived from radiometry and photometry would be strongly model dependent, there are some interesting special causes where this problem largely disappears. One simplification occurs if the thermal inertia of the surface is low; that is, if the illuminated face is at a temperature nearly in equilibrium with the insolation, while the dark hemisphere is much colder. In this case, nearly all of the absorbed sunlight is radiated from the sun-facing (and therefore Earth-facing, for a superior planet) hemisphere, which is the hemisphere we observe. Low thermal inertias appear to be ubiquitous among airless solar system bodies, as was noted in the previous section, presumably as the result of vacuum welding of fine regolith material produced from meteoric impacts. For the asteroids, theoretical models suggest that an insulating regolith should be maintained for objects down to a few kilometers in diameter; the only major exception might be for a largely metallic asteroid (Housen et al. 1979). A second simplification occurs for dark objects. In general, to interpret the radiometry one needs to know the relationship between Bond albedo and geometric albedo, but for a very dark object this information is not required. In effect, if an object is nearly black we can assume nearly all the incident sunlight goes to heating it, and the details of its photometric behavior are not important. Since most asteroids and many satellites are exceedingly dark, this effect considerably enhances the accuracy of the radiometric/photometric method.

In the satellite systems of the outer planets, infrared radiometry has been used to determine the diameters and albedos of Iapetus and several of the inner satellites of Saturn (Murphy et al. 1972; Morrison 1974b), but these results suffer from the fact that the albedos are not low, and in any case they will be superseded by Voyager results before this book is published. Of more significance is the measurement by Cruikshank (1979) of Hyperion, for which he obtained an albedo of 0.47 $\pm$ 0.11 and a diameter of 224 $\pm$ 30 km. These values are less likely to be improved upon by Voyager. In the Jovian system, Cruikshank (1977) has used this method to make the only measurements of Himalia (J6) and Elara (J7), for both of which he finds albedos of 0.03. His diameters are 170 $\pm$ 20 km for Himalia and 80 $\pm$ 20 km for Elara. Cruikshank et al. (1979) used an upper limit of the 20-μm flux of Triton to set an upper limit of 5200 km to the diameter and a lower limit to the geometric albedo of 0.19. This is an important result, inasmuch as these are the only data that tell us that Triton is not as large as Titan. The other outer planet satellites are undetectable with present instruments.

Asteroids have provided the most productive application of the radiometric/photometric technique. There are hundreds of asteroids large enough to have angular diameters of 0.1 arcsec as seen from Earth--too small to be resolved optically, of course, but large enough to produce a strong infrared signal. At the distance from the sun of the main asteroid belt, the temperatures result in peak thermal radiation in the 10 and 20-μm bands, where it is easily observed. In fact, if one were to look at the sky with 10-μm eyes, asteroids would provide a high percentage of the sources--as will become perhaps too apparent when IRAS begins its all-sky survey.

Only two alternative methods have been used to measure asteroid sizes and albedos. Stellar occultations observed photoelectrically from a network of sites provide by far the best values, but predictable events are rare and the area of the Earth's surface from which the observations can be made is small: comparable to that from which a total solar eclipse is visible. Half a dozen occultations have been observed, but the only two that were really successful were of 2 Pallas and 4 Juno. The results for these objects provide the best calibration for the radiometric/photometric technique. The other method is based on optical polarimetry at a variety of wavelengths. It has been applied to more than 50 asteroids, but it requires observations at several phase angles and it becomes unreliable for very dark objects. Most of the known asteroid diameters today were obtained from infrared observations; Morrison (1977) lists values for 187 objects. Most of these are estimated to be accurate to $\pm$10% in the diameter and $\pm$20% in the albedo. In the best observed cases, the accuracy may reach $\pm$5% in diameter, but disparities in the calibration based on the Juno and Pallas occultation results suggest an inherent uncertainty of at least ±5%.

All known asteroid diameters and albedos are listed along with other physical observations in the TRIAD (Tucson Revised Index of Asteroid Data) file, which was published in 1979 as a series of appendices to the University of Arizona volume *Asteroids*. The coordinated collection and interpretation of asteroid observations as represented in TRIAD has played a central role during the past decade in expanding our understanding of these objects. The first such effort to analyze spectral, polarimetric, and radiometric data for a statistically significant group of asteroids was made by Chapman et al. (1975) at a time when radiometric diameters were available for only 47 objects. Even then, it was apparent that most asteroids had extremely low albedos (<6%) and were chemically primitive. Chapman et al. designated these "C-class" asteroids by analogy with the carbonaceous chondrites; the next most populous class, which they called S objects, have higher albedos and spectra that frequently show evidence of iron-magnesium-silicate mineral assemblages. The classification of asteroids and its interpretation are discussed in detail by Bowell et al. (1978), while the actual mineralogy of these objects as derived from infrared reflectance spectroscopy is discussed by McCord and Cruikshank in this book.

The TRIAD file, which included 195 radiometric diameters when published in 1979, has been used for several statistical studies that provide the first reliable picture of the physical nature and distribution of the minor planets (Zellner and Bowell 1977; Zellner 1979; Chapman 1979). We now know what the numbers and distributions over size and distance from the Sun are for asteroids in each of several compositional classes. It is also possible to determine the compositional classes of objects related dynamically and presumed to have a common history. There is no room here to discuss these results, but it is enough to note that thermal radiometry is playing a central role in this fast-moving field as the primary means available today to measure the size of an asteroid and, perhaps even more important, its compositionally sensitive surface albedo. As a single specific example of these results, Table 2 lists the sizes and classifications of all asteroids 200 km or larger in diameter; before the widespread application of the radiometric/photometric technique to the asteroids, such a compilation would have been unthinkable.

## V. VOLCANIC ACTIVITY ON IO

In 1979, Voyager discovered that Io is a planetary object of incredible geologic activity, with eruptive plumes rising hundreds of kilometers above the surface, changes in albedo and color within a few weeks that affect areas of thousands of square kilometers, and localized hot spots with temperatures as much as 500 K above ambient

(Morabito et al. 1979; Smith et al. 1979a,b; Hanel et al. 1979). This unique level of volcanism is the outward manifestation of a molten interior, maintained by a major internal heat source. The most probable cause of the internal heating is tidal stressing of Io that results from its non-circular orbit, which is in turn dynamically coupled to the other Galilean satellites. Initial calculations of the magnitude of the internal heat source suggested values for the power dissipated as high as $10^{13}$ watts (Peale et al. 1979). Luminosity of this scale can be measured with Earth-based infrared techniques.

TABLE 2

THE LARGEST ASTEROIDS

| Asteroid | Type | Diam. (km) | Asteroid | Type | Diam. (km) |
|---|---|---|---|---|---|
| 1 Ceres | C | 1025 | 24 Themis | C | 249 |
| 2 Pallas | U | 583 | 3 Juno | S | 249 |
| 4 Vesta | U | 555 | 16 Psyche | M | 249 |
| 10 Hygeia | C | 443 | 13 Egeria | C | 245 |
| 704 Interamnia | U | 338 | 216 Kleopatra | CMEU | 236? |
| 511 Davida | C | 335 | 165 Loreley | C | 228 |
| 65 Cybele | C | 311 | 19 Fortuna | C | 226 |
| 52 Europa | C | 291 | 7 Iris | S | 222 |
| 451 Patientia | C | 281 | 532 Herculina | S | 219 |
| 31 Euphrosyne | C | 270 | 250 Bettina | CMEU | 211? |
| 15 Eunomia | S | 261 | 702 Alauda | CU | 217 |
| 324 Bamberga | C | 256 | 747 Winchester | C | 208 |
| 107 Camilla | C | 252 | 432 Diotima | C | 209 |
| 87 Sylvia | CMEU | 251? | 386 Siegena | C | 203 |
| 45 Eugenia | U | 250 | 375 Ursula | C | 200 |

With the clear perspective of hindsight, it is now clear that infrared observers had detected the effects of Ionian volcanism long before Voyager. In 1973, 10-μm photometry of Io obtained with the Hale 5-meter telescope during eclipses showed a remarkably high flux density, quite different from the cooling curves of the other Galilean satellites (Hansen 1973). At the same time, 20-μm eclipse observations with the Mauna Kea 2.2-meter telescope displayed more normal behavior (Morrison and Cruikshank, 1973b). It is now clear that almost all of the 10-μm radiation seen by Hansen originated in spots at temperatures of 200 K or higher covering a small fraction of the surface. A more direct measurement of a thermal outburst on Io was made in 1978 by Witteborn et al. (1979), who observed a short-lived but dramatic enhancement at 5 μm; unfortunately, they considered but then rejected the volcanic hypothesis. Between the two Voyager flybys Sinton (1980) observed another similar 5-μm event, but by then the correct explanation had become dramatically apparent.

Ground-based observations are capable of detecting some form of volcanic outbursts on Io, but these events are rare and it is not at all clear just what physical mechanisms are involved. More important is the ability of infrared astronomy to measure the total emitted power from Io. If a source of magnitude $10^{13}$ watts resulted in a uniform heat flow through the crust, such a measurement would not be possible, since the total rise in surface temperature would be only about 1 K. However, we are fortunate that the escape of energy from Io takes a more readily measurable route. Voyager 1 (Hanel et al. 1979) measured discrete hot spots, with temperatures hundreds of degrees above the background, and it is similar hot spots that contribute to the anomalous thermal behavior of Io. The elevation of the 10-μm temperature during eclipse has already been noted, and even outside of eclipses it is clear that the brightness temperature of Io increases markedly toward shorter wavelengths. Matson et al. (1980) have analyzed these spectral data in terms of a hot-spot model and calculated that $2 \pm 1$ watts $m^{-2}$ was being released through hot spots, corresponding to an internal power of about $10^{14}$ watts.

The effects of the hot spots are even more dramatic at shorter wavelengths when Io is in eclipse, as discussed by Sinton et al. (1980): even at wavelengths as short as 2 μm the satellite remains detectable, indicating that some areas on the surface have temperatures as high as 600 K. Eclipse photometry covering the entire range from 3 to 30 μm was obtained in April 1980 by Morrison and Telesco (1981). These measurements of the hot spots yield an average heat flow of $1.5 \pm 0.5$ watts $m^{-2}$ or an internal power of $(6 \pm 2) \times 10^{13}$ watts. The models fit to the spectrum suggest that there is a broad range of temperatures, from about 500 K (covering a few millionths of the surface) down to about 200 K (covering nearly 1% of the surface).

The observational determinations of the Ionian heat flow as represented by the hot spots are summarized in Table 3, together with the results of dynamical calculations. It is important to remember that the observational values refer only to areas with temperatures substantially above the mean; the temperature increase due to a more uniform heat flow could not be detected. Thus the numbers given represent a lower limit to the internal energy source, if we adopt the assumptions that the recent observations of the Jupiter-facing hemisphere of Io are representative of the entire surface over long periods of time. This lower limit is in excess of $10^{13}$ watts, and may be as high as $10^{14}$ watts. Yoder (1979,1980) has argued that the dynamics of the Jovian satellite system do not permit the deposition of more than a few times $10^{13}$ watts of tidal energy in Io on an equilibrium basis. Clearly, the infrared measurements not only are fascinating for the information they provide us on volcanic processes; in addition, they appear to challenge the basic mechanics of origin of the internal heat of Io. These topics are pursued in detail in chapters by Greenberg (1981), Cassen et al. (1981), and Pearl and Sinton (1981) in the forthcoming book *The Satellites of Jupiter*.

TABLE 3

VALUES FOR THE INTERNAL HEAT SOURCE OF IO

| Method | Luminosity (watts) | Heat Flow (W $m^{-2}$) | Reference |
|---|---|---|---|
| Theory: | $\lesssim 4 \times 10^{12}$ | $\lesssim 0.1$ | Peale et al. (1979) |
| Theory: | $10^{12} - 10^{13}$ | 0.02 - 0.2 | Yoder (1979) |
| Theory: | $2 \times 10^{12} - 4 \times 10^{13}$ | 0.05 - 1.0 | Yoder (1980) |
| Spectrum: | $(8 \pm 4) \times 10^{13}$ | $2 \pm 1$ | Matson et al. (1980) |
| Voyager: | $\sim 10^{14}$ | ~2 | Pearl (1980) |
| Eclipse: | $(7 \pm 3) \times 10^{13}$ | $1.8 \pm 0.8$ | Sinton (1981) |
| Eclipse: | $(6 \pm 2) \times 10^{13}$ | $1.5 \pm 0.5$ | Morrison & Telesco (1981) |

## VI. CONCLUSIONS

As the foregoing examples indicate, thermal infrared studies of planets and smaller solar system bodies are used to investigate a variety of scientific problems. In a few cases, such as determining the nature of the phase changes on the sublimating Martian polar cap, the actual temperatures are of direct interest. However, most of these investigations have as their goal the study of the more fundamental physical nature of planetary surfaces and interiors, and of the processes influencing them. It seems certain that such studies, carried out from the ground, from high in the terrestrial atmosphere, from Earth orbit, and from planetary flybys and orbiters, will continue to play an important part in efforts to explore and understand the planets and their origins.

For their advice and assistance I thank H. H. Kieffer, T. Z. Martin, F. D. Palluconi, and W. M. Sinton, and I especially am grateful to D. P. Cruikshank for his encouragement and aid in the preparation of this chapter. This research was supported in part by NASA Grants NGL 12-001-057 and NSG 7633.

REFERENCES

Allen, D.A.: 1970, Nature 227, pp. 158-159.
Aumann, H.H., Gillespie, C.M., and Low, F.J.: 1969, Astrophys. J. 157, pp. L69-L72.
Bowell, E., Chapman, C.R., Gradie, J.C., Morrison, D., and Zellner, B.: 1978, Icarus 35, pp. 313-335.
Bodenheimer, P.: 1974, Icarus 23, pp. 319-325.
Cassen, P.M., Reynolds, F.T., and Peale, S.J.: 1981, in D. Morrison (ed.), "The Satellites of Jupiter," Univ. of Arizona Press, Tucson (in press).
Chapman, C.R.: 1979, in T. Gehrels (ed.), "Asteroids," Univ. of Arizona Press, Tucson, pp. 25-60.
Chapman, C.R., Morrison, D., and Zellner, B.: 1975, Icarus 25, pp. 104-130.
Chase, S.C., Miner, E.D., Morrison, D., Münch, G., and Neugebauer, G.: 1976, Icarus 28, pp. 565-578.
Chase, S.C., Ruiz, R.D., Münch, G., Neugebauer, G., Schroeder, M., and Trafton, L.M.: 1974, Science 183, pp. 315-317.
Courtin, R., Lena, P., de Muizon, M., Rouan, D., Nicollier, C., and Wijnbergen, J.: 1979, Icarus 38, pp. 411-419.
Cruikshank, D. P.: 1977, Icarus 30, pp. 224-230.
Cruikshank, D.P.: 1979, Icarus 37, pp. 307-309.
Cruikshank, D.P., Stockton, A., Dyck, H.M., Becklin, E.E., and Macy, W.: 1979, Icarus 40, pp. 104-114.
Cuzzi, J.: 1978, in D.M. Hunten and D. Morrison (eds.), "The Saturn System," NASA CP-2068, pp. 73-104.
Erickson, E.F., Goorvitch, D., Simpson, J.P., and Strecker, D.W.: 1978, Icarus 35, pp. 61-73.
Gatley, I., Kieffer, H., Miner, E., and Neugebauer, G.: 1974, Astrophys. J. 190, pp. 497-503.
Graboske, H.C., Pollack, J.B., Grossman, A.S., and Olness, R.J.: 1975, Astrophys. J. 199, pp. 265-281.
Greenberg. R.: 1981, in D. Morrison (ed.), "The Satellites of Jupiter," Univ. of Arizona Press, Tucson (in press).
Hanel, R., and the Voyager IRIS Team: 1979, Science 204, pp. 972-976.
Hansen, O.L.: 1973, Icarus 18, pp. 237-246.
Housen, K.R., Wilkening, L.L., Chapman, C.R., and Greenberg, R.J.: 1979, in T. Gehrels (ed.), "Asteroids," Univ. of Arizona Press, Tucson, pp. 601-627.
Ingersoll, A.P., Münch, G., Neugebauer, G., Diner, D.J., Orton, G.S., Schupler, B., Schroeder, M., Chase, S.C., Ruiz, R.D., and Trafton, L.M.: 1975a, Science 188, pp. 472-473.
Ingersoll, A.P., Münch, G., Neugebauer, G., and Orton, G.S.: 1975b, in T. Gehrels (ed.), "Jupiter," Univ. of Arizona Press, Tucson, pp. 197-205.
Ingersoll, A.P., Orton, G.S., Münch, G., Neugebauer, G., and Chase, S.C.: 1980, Science 207, pp. 439-443.
Jones, T.J., and Morrison, D.: 1974, Astron. J. 79, pp. 892-895.
Kieffer, H.H., Chase, S.C., Miner, E., Münch, G., and Neugebauer, G.: 1973, J. Geophys. Res. 78, pp. 4291-4312.

Kieffer, H.H., Martin, T.Z., Peterfreund, A.R., Jakosky, B.M., Miner, E.D., and Palluconi, F.D.: 1977, J. Geophys. Res. 82, pp. 4249-4291.
Loewenstein, R.F., Harper, D.A., Moseley, S.H., Telesco, C.M., Thronson, H.A., Hildebrand, R.H., Whitcomb, S.E., Winston, R., and Stiening, R.F.: 1977a. Icarus 31, pp. 315-324.
Loewenstein, R.F., Harper, D.A., and Moseley, S. H.: 1977b, Astrophys. J. 218, pp. L145-L146.
Low, F.J.: 1966, Astron. J. 71, p. 391 (abstract).
Matson, D.L.: 1971, in T. Gehrels (ed.), "Physical Studies of Minor Planets," NASA SP-267, pp. 45-50.
Matson, D.L., Veeder, G.J., and Lebofsky, L.A.: 1978, in D. Morrison and W.C. Wells (eds.), "Asteroids: An Exploration Assessment," NASA CP-2053, pp. 127-144.
Matson, D.L., Ransford, G.A., and Johnson, T.V.: 1980, J. Geophys. Res. (in press).
Morabito, L.A., Synnott, S.P., Kupferman, P.N., and Collins, S.A.: 1979, Science 204, p. 972.
Morrison, D.: 1968, Smithsonian Astrophys. Obs. Special Report #284.
Morrison, D.: 1973, Icarus 19, pp. 1-14.
Morrison, D.: 1974a, Icarus 22, pp. 57-64.
Morrison, D.: 1974b, Icarus 22, pp. 51-56.
Morrison, D.: 1977, Icarus 31, pp. 185-220.
Morrison, D., and Cruikshank, D.P.: 1973a, Astrophys. J. 179, pp. 329-331.
Morrison, D., and Cruikshank, D.P.: 1973b, Icarus 18, pp. 224-236.
Morrison, D., and Lebofsky, L.A.: 1979, in T. Gehrels (ed.), "Asteroids," Univ. of Arizona Press, Tucson, pp. 184-205.
Morrison, D., and Telesco, C.: 1981, Icarus, submitted.
Murphy, R.E., Cruikshank, D.P., and Morrison, D.: 1972, Astrophys. J. 177, pp. L93-L96.
Murray, B.C., Westphal, J.A., and Wildey, R.L.: 1965, Astrophys. J. 141, pp. 1590-1592.
Neugebauer, G., Münch, G., Kieffer, H.H., Chase, S.C., and Miner, E.: 1971, Astron. J. 76, pp. 719-728.
Palluconi, F.D., and Kieffer, H.H.: 1981, Icarus (in press).
Peale, S.J., Cassen, P.M., and Reynolds, R.T.: 1979, Science 203, pp. 892-894.
Pearl, J.C.: 1980, paper given at IAU Colloquium No. 57.
Pearl, J.C., and Sinton, W.M.: 1981, in D. Morrison (ed.), "The Satellites of Jupiter," Univ. of Arizona Press, Tucson (in press).
Pettit, E.: 1961 in G.P. Kuiper and B.M. Middlehurst (eds.), "Planets and Satellites," Univ. of Chicago Press, Chicago, pp. 400-428.
Pollack, J.B.: 1978, in D. M. Hunten and D. Morrison (eds.), "The Saturn System," NASA CP-2068, pp. 9-28.
Pollack, J.B., Grossman, A.S., Moore, R., and Graboske, H.C.: 1977, Icarus 30, pp. 111-128.
Rieke, G.H., and Low, F.J.: 1974, Astrophys. J. 193, pp. L147-L148.
Stier, M.T., Traub, W.A., Fazio, G.G., Wright, E.L., and Low, F.J.: 1978, Astrophys. J. 226, pp. 347-349.

Sinton, W.M.: 1961, in G.P. Kuiper and B.M. Middlehurst (eds.,) "Planets and Satellites," Univ. of Chicago Press, Chicago, pp. 429-441.
Sinton, W.M.: 1980, Astrophys. J. 235, pp. L49-L51.
Sinton, W.M.: 1981, J. Geophys. Res. (in press).
Sinton, W.M., and Strong, J.: 1960, Astrophys. J. 131, pp. 459-469.
Sinton, W.M., Tokunaga, A., Becklin, E.E., Gatley, I., Lee, T.J., and Lonsdale, C.: 1980, Science (in press).
Smith, B.A. and the Voyager Imaging Team: 1979a, Science 204, pp. 951-972.
Smith, B.A. and the Voyager Imaging Team: 1979b, Science 206, pp. 927-950.
Trafton, L.: 1974, Astrophys. J. 193, pp. 477-480.
Witteborn, F.C., Bregman, J.D., and Pollack. J.B.: 1979, Science 203, pp. 643-646.
Yoder, C.F.: 1979, Nature 279, pp. 767-770.
Yoder, C.F.: 1980, paper given at IAU Colloquium No. 57.
Zellner, B.: 1979, in T. Gehrels (ed.), "Asteroids," Univ. of Arizona Press, Tucson, pp. 783-808.
Zellner, B., and Bowell, E.: 1977, in A. H. Delsemme (ed.), "Comets, Asteroids, and Meteorites," Univ. of Toledo, Toledo, pp. 185-197.
Zimbelman, J.R., and Kieffer, H.H.: 1979, J. Geophys. Res. 84, pp. 8239-8251.

DISCUSSION FOLLOWING PAPER DELIVERED BY D. MORRISON

BEER: Do the inferences about the compactness of the surface of Callisto support the old ideas about the structure of regoliths _in vacuo_? That is, can we expect a "fairy castle" structure?

MORRISON: The inference about the compactness comes from the modeled thermal conductivity of the surfaces of the satellites. The thermal conductivity of Callisto appears to be within a factor of two or three of that of the Moon, and I believe we can assume that the surface structure is therefore similar to that of the Moon. You will recall that the Moon has very low thermal conductivity and a rather compact surface, and that the surface has substantial bearing strength.

WERNER: How do your inferences from the ground-based measurements of the surface properties of the Jovian satellites compare with the interpretation of the Voyager photographs?

MORRISON: At present these two data sets appear to be orthogonal. The disparity between our thermal measurements of surfaces on a length scale of a few centimeters and the maximum Voyager resolution of about 1 kilometer is very great. Perhaps we can come closer by looking at the results of the Viking measurements of Mars where we have a lander that can help connect the two data sets. In any case, I think there is room for a great deal of interpretation in the analysis of these data.

ALLEN: A decade ago when infrared measurements of asteroid diameters and albedos were first made I felt that one would never achieve better than 10% accuracy on diameters--hence 20% on albedos--due to the effects of the irregular surface of the asteroids and uncertainties in the amount of energy radiated from the dark side. Thus it doesn't surprise me that the accuracy you are attaining compared to occultation measures is 10%. Now, I suspect that these effects (especially shape) will increase with smaller asteroids. Now I see that you are pushing down below 50 km diameters, and I'd like to ask how small you think you can go before the errors become too large to be useful.

MORRISON: This is an excellent question. Of course, 10% accuracy in diameter and 20% in albedo, or even worse, is quite enough for the broad compositional classification, so our interest in pushing for higher accuracy is either for aesthetic reasons or interest in determinations of the densities of asteroids. If, however, the regolith of an asteroid is completely removed, a much higher fraction of the incident solar radiation is emitted from the anti-solar side, so that the computational scheme used so far in this work is no longer completely valid. Calculations indicate that the non-regolith limit might be on the order of 10 km diameter, so we might be able to see the effects on the small Earth-approaching asteroids. There is some evidence that a few asteroids of small size, particularly Betulia, emit substantially less infrared thermal radiation than expected. That is, their diameters determined by the radiometric technique are substantially smaller than those found by other means. These may represent cases of small asteroids without regoliths. For some small asteroids the radiometric technique seems to work, however.

JOSEPH: Why is gravitational contraction adequate to account for the excess power radiated by Jupiter (over that absorbed from the Sun), but not for Saturn?

MORRISON: I have not personally made the calculations, but Pollack, Bodenheimer, and others predict a weaker internal heat source for Saturn than for Jupiter. My intuitive feeling is that because Saturn is a smaller object we should expect less initial heating per unit mass as it collapses gravitationally, whereas the observations indicate a greater power per unit mass for Saturn. Thus an additional energy source is indicated.