Tectonics of the outer planet satellites

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**Summary**

Tectonic features on the satellites of the outer planets range from the familiar, such as clearly recognizable graben on many satellites, to the bizarre, such as the ubiquitous double ridges on Europa, the twisting sets of ridges on Triton, or the isolated giant mountains rising from Io’s surface. All of the large and middle-sized outer planet satellites except Io are dominated by water ice near their surfaces. Though ice is a brittle material at the cold temperatures found in the outer solar system, the amount of energy it takes to bring it close to its melting point is lower than for a rocky body. Therefore, some unique features of icy satellite tectonics may be influenced by a near-surface ductile layer beneath the brittle surface material, and several of the icy satellites may possess subsurface oceans. Sources of stress to drive
tectonism are commonly dominated by the tides that deform these satellites as they orbit their primary giant planets. On several satellites, the observed tectonic features may be the result of changes in their tidal figures, or motions of their solid surfaces with respect to their tidal figures. Other driving mechanisms for tectonics include volume changes due to ice or water phase changes in the interior, thermoelastic stress, deformation of the surface above rising diapirs of warm ice, and motion of subsurface material toward large impact basins as they fill in and relax. Most satellites exhibit evidence for extensional deformation, and some exhibit strike-slip faulting, whereas contractional tectonism appears to be rare. Io’s surface is unique, exhibiting huge isolated mountains that may be blocks of crust tilting and foundering into the rapidly emptying interior as the surface is constantly buried by deposits from hyperactive volcanoes. Of the satellites, diminutive Enceladus is spectacularly active; its south polar terrain is a site of young tectonism, copious heat flow, and tall plumes venting into space. Europa’s surface is pervasively tectonized, covered with a diverse array of exotic and incompletely understood tectonic features. The paucity of impact craters on Europa suggests that its tectonic activity is ongoing. Geysers on Triton show that some degree of current activity, while tectonic features that cross sparsely cratered terrain indicate that it may also be tectonically active. Ganymede and Miranda both exhibit ancient terrains that have been pulled apart by normal faulting. On Ganymede these faults form a global network, while they are confined to regional provinces on Miranda. Ariel, Dione, Tethys, Rhea, and Titania all have systems of normal faults cutting across their surfaces, though the rifting is less pronounced than it is on Ganymede and Miranda. Iapetus exhibits a giant equatorial ridge that has defied simple explanation. The rest of the large and middle-sized satellites show very little evidence for tectonic features on their surfaces, though the exploration of Titan’s surface has just begun.

1 Introduction

The four new worlds orbiting Jupiter, as reported in Galileo’s Siderius Nuncius (Galilei, 1610), forever changed humankind’s worldview by demonstrating that Jupiter, like Earth, is a center of celestial motion, in strong support of the Copernican model of the heavens (Copernicus, 1543). Four centuries later, the consequences of this seemingly simple motion of satellites about their primary planet are still being understood. On the worlds now known as the Galilean satellites of Jupiter, as well as the satellites of the other giant planets, Saturn, Uranus, and Neptune, the interactions of tides with satellite interiors has played a large part in determining geological activity. Tides raised by the giant planets power volcanoes on Io, maintain a liquid water ocean beneath Europa’s ice, drive plumes on Enceladus, and probably played a large role in fracturing the surfaces of Ganymede, Dione, and other satellites.
We begin this chapter by examining the behavior of ice lithospheres, contrasting the rheology of ice to the more familiar rock behavior on the inner planets. We then summarize the global and local stress mechanisms that can affect outer planet satellites, as linked to their geophysics and geodynamics. After this background, we examine the tectonics on groups of outer planet satellites, starting with Io, the only large rocky outer planet satellite, then proceeding from currently active icy satellites, to satellites that formerly had tectonic activity, to satellites with very little evidence of any activity. We rely on insights gained from analysis of data from the twin Voyager spacecraft, which flew through all of the giant planet systems between 1979 and 1989; from the Galileo spacecraft, which orbited Jupiter from 1995 to 2003; and from the Cassini mission, which arrived in orbit around Saturn in 2004. The chapters that comprise Burns and Matthews (1986), Morrison (1982), Gehrels and Matthews (1984), Bergstralh et al. (1991), and Cruikshank (1995) offer excellent reviews of Voyager-based understanding of the satellites. Comprehensive Galileo-based syntheses of Jupiter’s satellites are provided by several chapters of Bagenal et al. (2004), and the understanding of Saturn’s satellites from Cassini is currently evolving as the spacecraft mission proceeds.

2 Rheology of ice

2.1 Introduction

The observable consequences of tectonic stresses depend mainly on the response of the material being stressed, i.e., its rheology. Hence, understanding the rheology of ice and rock is of fundamental importance to interpreting the tectonics of outer solar system satellites. In this section, we focus on the rheology of water ice, because its behavior is less well known to terrestrial geologists and it has some important differences to the behavior of rock. In particular, we discuss the ways in which ice may respond to imposed stresses, and the consequences of these different response mechanisms.

It is well known that a material’s rheological properties depend on its homologous temperature, that is the ratio of its absolute temperature to the melting temperature (Frost and Ashby, 1982). On Earth, ice is rarely at temperatures lower than about 80% of the melting point (a homologous temperature of 0.8). On the icy satellites, typical surface temperatures correspond to a homologous temperature of about 0.4, similar to that of rocks at the surface of the Earth. Thus, one would expect the behavior of ice at the surface of the icy satellites to mimic that of rocks on Earth, and this is exactly what is observed; as discussed later in this chapter, many tectonic features observed on icy satellites have counterparts on the Earth.
As we also discuss further below, water ice has several important differences when compared to silicate materials. First of all, solid ice is less dense than its molten equivalent (water). This makes the surface eruption of water difficult, and means that any density-driven overturn (e.g. solid state convection or subduction) would have to take place entirely within a floating ice shell. In contrast to most terrestrial contexts, a modest thermal gradient can allow ice to reach a high homologous temperature at relatively shallow depths (several km). Ice close to its melting point flows more readily than silicates. Thus, viscous flow timescales are much shorter in icy bodies than in their silicate equivalents (Section 2.4), and a ductile layer may be more shallow, with more influence on the surface, than in a typical terrestrial context. In addition to being more ductile, ice is intrinsically weaker than rock in that it is less rigid (lower Young’s modulus, Section 2.2), and undergoes brittle tensile and compressive failure at lower stresses (Weeks and Cox, 1984).

As with silicate materials, ice under stress can respond in one of three idealized ways. At low stresses and strains, the ice will deform in an elastic (recoverable) manner, but at strains greater than roughly $10^{-4}$, the ice will undergo irrecoverable
deformation. At low temperatures and relatively high strain rates, this deformation will be accomplished by brittle failure. At higher temperatures and lower strain rates, the result will be ductile behavior or creep. A closer approximation to reality is to say that materials behave in an elastoviscoplastic manner, combining elements of elastic and ductile behavior, as well as brittle or “plastic” failure.

2.2 Elastic deformation
At low stresses and strains, ice will deform elastically, and the relationship between stress $\sigma$ and strain $\varepsilon$ depends on the Young’s modulus $E$ of the material as $\sigma = E\varepsilon$ (for uniaxial deformation). Measurements of Young’s modulus in small laboratory specimens of ice are straightforward and yield a value near 9 GPa (Gammon et al., 1983). However, the effective Young’s modulus of large bodies of deformed ice is less obvious (e.g., Nimmo, 2004a). Observations of ice shelf response to tides on Earth (Vaughan, 1995) give an effective $E \sim 0.9$ GPa, an order of magnitude smaller than the laboratory values. This discrepancy is most likely due to the fact that a large fraction of the ice shelf thickness is not responding in a purely elastic fashion (e.g., Schmeltz et al., 2002). On icy satellites, porosity and/or fracturing in the near-surface may result in a reduction in the effective value of $E$ (Nimmo and Schenk, 2006).

2.3 Brittle deformation
At shallow depths, where the overburden pressure is low, ice can undergo tensile failure (for a review of ice fracture, see Schulson, 2006). The tensile strength of ice at temperatures above $-50^\circ$C has been studied for glaciological applications, but preliminary experiments of tensile failure of polycrystalline ice at much colder temperatures relevant to icy satellite surfaces (Sklar et al., 2008) show that tensile strength rises with decreasing temperature, up to a few MPa. At greater depths, the overburden pressures are such that failure occurs by shear motion. For silicate materials, shear failure on preexisting faults occurs when the shear stresses exceed some fraction, typically 0.85, of the normal stresses (e.g., Turcotte and Schubert, 2002). This behavior is largely independent of composition and is known as Byerlee’s rule. Cold ice in the laboratory obeys a Byerlee-like rule at low sliding velocities and stresses, with a coefficient of friction $\mu_f \approx 0.55$ (Beeman et al., 1988). At higher sliding velocities, the behavior becomes more complex (Rist, 1997), with ice friction dependent on temperature at sufficiently high sliding velocities (Petrenko and Whitworth, 1999; Maeno and Arakawa, 2004). Brittle deformation is expected to dominate on icy satellites at shallow depths where normal stresses are small and temperatures are low.
2.4 Ductile deformation

At sufficiently high temperatures, ice responds to applied stress by deforming in a ductile fashion, also known as creep. The response is complicated by the fact that individual ice crystals can deform in several different ways: by diffusion of defects within grain interiors, by sliding of grain boundaries, and by dislocation creep (Goldsby and Kohlstedt, 2001; Durham et al., 2001). Which mechanism dominates depends on the specific stress and temperature conditions, but each individual mechanism can be described by the following generalized equation:

$$\dot{\varepsilon} = A\sigma^n d^{-p} \exp \left( -\frac{Q + PV}{RT} \right).$$

(7.1)

Here $\dot{\varepsilon}$ is the resulting strain rate of the deforming ice, $\sigma$ is the differential applied stress, $A$, $n$ and $p$ are experimentally determined constants, $d$ is the grain size, $Q$ and $V$ are the experimentally determined activation energy and volume, respectively, $R$ is the gas constant, and $P$ and $T$ are pressure and temperature, respectively. Here the strain rate and stress variables are scalar representations of the appropriate tensors. For icy satellites, the $PV$ contribution is generally small enough to be ignored, and strain rates increase with increasing temperature and stress and decreasing grain size, as expected. Because several different deformation mechanisms can operate together, the total strain rate is a function of the individual strain rates (see Equation 3 in Goldsby and Kohlstedt, 2001).

For a given ice grain size, the applied stress and temperature control the deformation mechanism (see Figure 1 in Barr and Pappalardo, 2005). At low stresses and high temperatures, diffusion creep is expected to dominate and is predicted to result in Newtonian flow (that is, $n = 1$, a linear relationship between stress and strain rate) with a grain-size dependence ($p = 2$). At higher stresses and lower temperatures, the dominant creep regimes are basal slip and grain boundary sliding, which result in non-Newtonian behavior ($n \sim 2$) and grain-size dependence. At even higher stresses, strongly non-Newtonian dislocation creep ($n = 4$) dominates. Deformation rates are enhanced within about 20 K of the melting temperature (Goldsby and Kohlstedt, 2001), presumably because of the presence of thin films of water along grain boundaries (e.g., De La Chapelle et al., 1999).

Because stresses and to some extent strain rates within the warm interiors of icy satellites are expected to be low, the most relevant deformation mechanism within the warm icy interior is probably diffusion creep (McKinnon, 2006; Moore, 2006), while other mechanisms may be important at higher stresses (for instance, near the surface). Diffusion creep has the advantage of resulting in Newtonian behavior, but the disadvantage that the viscosity ($\eta \approx \sigma/\dot{\varepsilon}$) is dependent on the grain size. Ice grain size evolution is poorly constrained, because it depends both on the presence
of secondary (pinning) phases and because of dynamic recrystallization processes (e.g., Barr and McKinnon, 2007). Given the uncertainties, it is often acceptable to assume for modeling purposes that ice has a Newtonian viscosity near its melting temperature in the range $10^{13} - 10^{15}$ Pa s (e.g., Pappalardo et al., 1998a). However, more complex models taking into account the non-Newtonian and viscoelastic behavior of ice have also been developed (e.g., Dombard and McKinnon, 2006a), defining an effective viscosity assuming an effective composite strain rate.

Although grain size is a major unknown for describing the ductile deformation of ice in outer planet satellites, other effects can also be important. The presence of even small amounts of fluid significantly enhances creep rates (e.g., De La Chapelle et al., 1999). On the other hand, the presence of rigid impurities (e.g., silicates) at levels greater than 10 percent serves to increase the viscosity (Friedson and Stevenson, 1983; Durham et al., 1992). Finally, some higher pressure phases of ice, clathrates incorporating other chemical species such as methane, and ice with hydrated sulfate salts, all tend to have much higher viscosities than pure water ice at the same $P, T$ conditions (Durham et al., 1998, 2005; McCarthy et al., 2007), whereas ammonia–water ice mixtures and other ice phases can be weaker (Durham et al., 1993; Durham and Stern, 2001).

### 2.5 Viscoelastic behavior

In reality, materials do not exhibit entirely elastic or entirely viscous behavior. Rather, they exhibit elastic-like behavior if the timescale over which deformation occurs is very short compared to a characteristic deformation timescale of the material (proportional to $\eta/E$), known as the Maxwell time $\tau_M$ (Turcotte and Schubert, 2002). Conversely, if the deformation timescale is long compared to the Maxwell time, the material behaves in a viscous fashion. Such compound materials are termed viscoelastic. The principal importance of viscoelastic materials with reference to icy satellites is that the amount of tidal heating in an ice layer may be controlled by the viscoelastic properties of the ice, in particular its viscosity structure (e.g., Ross and Schubert, 1986).

### 2.6 Application to icy satellites

Temperatures near the surface of icy satellites are sufficiently cold, and overburden pressures sufficiently small, that tectonic stresses are likely to result in brittle deformation. However, at greater depths, temperatures will increase, allowing ductile deformation to dominate. At any particular depth, the mechanism with the lowest yield stress (ductile/elastic/brittle) dominates. For example, in pure extension, brittle failure at the surface transitions to ductile flow at depth. As another example, if
Figure 7.2. Schematic stress profile within generic icy satellite shell. This example shows how the mechanical layers interact in the case of bending the shell, with opposite stresses at the top and bottom. Near the surface the ice is cold and brittle and stress increases proportionately with overburden pressure. The elastic bending stresses decrease towards the midpoint of the shell and lead to an elastic “core”; the slope of the elastic stress curve depends on the local curvature of the shell and the Young’s modulus (e.g., Turcotte and Schubert, 2002). At greater depths, the ice is sufficiently warm that the shell deforms in a ductile fashion. The first moment of the stress profile about the midpoint controls the effective elastic thickness $T_e$ of the ice shell as a whole (e.g., Watts, 2001). Note that $T_e$ is dominated by the brittle and elastic portions of the shell (see Nimmo and Pappalardo, 2004).

The principal source of stress is bending (e.g., due to fault motion), then near the mid-plane of the bending shell the stresses are sufficiently low that deformation is accommodated elastically. Thus, one would expect an ice layer deformed in bending to consist of three regions (Figure 7.2): a brittle near-surface layer; an elastic “core”; and a ductile base (e.g., Watts, 2001). The upper and lower halves of the shell experience opposite stresses, hence the change in sign.

The thickness of the near-surface brittle layer in Figure 7.2 depends mainly on the temperature gradient, and to a lesser extent on the degree of curvature (bending). Thus, if the brittle layer thickness can be constrained, for example by observations of fault spacing (Jackson, 1989), then the temperature structure can be deduced for a given tectonic interpretation (e.g., Golombek and Banerdt, 1986).

It has long been recognized that the brittle/elastic/ductile structure encountered in real silicate or icy bodies can be more simply modeled as a purely elastic layer overlying an inviscid interior (e.g., McNutt, 1984). The thickness of this modeled elastic layer, termed the effective elastic thickness $T_e$, may in some cases be deduced from observations such as surface topography when a load is bending the elastic layer (e.g., Watts, 2001). Furthermore, the value of $T_e$ thus deduced may be related back to the more realistic stress profile shown in Figure 7.2 if the topographic curvature is known (McNutt, 1984; Watts, 2001). Because the stress
profile depends primarily on temperature gradient, determination of the effective elastic thickness of the lithosphere can be used along with strength envelopes such as that shown in Figure 7.2 to determine the thermal structure of ice shells (e.g., Nimmo and Pappalardo, 2004). For a conductive ice layer, knowing the thermal structure in turn allows the layer thickness to be deduced.

Many of the equations used in the rest of this chapter implicitly make the simplifying assumption that tectonic behavior on icy satellites is adequately modeled using the elastic lithosphere model. We emphasize here that $T_e$ is the thickness of the elastic layer which deforms in a manner equivalent to the more complicated strength envelope likely to be present in real satellites. The utility of $T_e$ is that it can be measured directly, and that it may be used to infer the real strength envelope and thus the thermal gradient. The drawback of this approach is that even cold ice topography may viscously relax over geologic time (e.g., Dombard and McKinnon, 2006a).

### 2.7 Comparison with silicate behavior

Much of the above analysis can be equally applied to silicate materials, such as the lithosphere of Io. However, there are some important differences. Silicate materials have higher Young’s moduli ($\sim 100$ GPa) and near their melting point have viscosities $\sim 10^4$–$10^6$ times higher than ice, implying much longer Maxwell timescales. Silicate materials have much higher melting temperatures ($\sim 950$ to 1500 K, or higher, depending on pressure; Best, 2003) than ices, so a given homologous temperature is achieved at much greater depth within a rocky body than an icy body for a given thermal gradient. The low relative density of silicate melts means that the melting products are much more likely to be erupted to the surface than is water on an icy body. Silicate melts are typically more viscous than water, which means melt drainage is slower and thus that partially molten regions are likely to persist for longer in silicate systems than in ices.

### 3 Global and local stress mechanisms

The tectonic features observed on a satellite provide clues to its geological and orbital history. In order to understand the origin of the strain represented by landforms on outer planet satellites, we must understand the possible range of stress mechanisms that may have operated on the satellites. Here we review processes likely to be most relevant, grouping them into global and local mechanisms. A seminal review of this material is provided by Squyres and Croft (1986). As before, our focus will be on icy satellites; where appropriate we will also mention mechanisms relevant to silicate bodies.
3.1 Background

3.1.1 Satellite figures

The global stresses experienced by satellites commonly relate to changes in their shape, so we present a brief discussion of satellite figures here. Satellite figures depart from a spherical shape primarily due to rotation and tides. The relative magnitude of the two effects is given by $\omega^2 a_s^3 / 3GM_p$ (Murray and Dermott, 1999), where $\omega = 2\pi / P_s$ is the rotational frequency of the satellite, $a_s$ is the semimajor axis of its orbit, $G$ is the gravitational constant, and $M_p$ is the mass of the primary planet. For satellites in synchronous rotation ($P_s =$ orbital period = rotational period), the tidal and rotational potentials that characterize the satellite shape are in the ratio 3:1. Thus the tidal bulge of a synchronously rotating satellite is larger than the rotational bulge along the equator.

The equilibrium shape of a satellite is the shape it would assume if it is in hydrostatic equilibrium, that is, if its shape approximates that of a fluid that is incapable of supporting global-scale forces by elastic (or dynamic) stresses. In this case, the satellite will have an ellipsoidal shape with three unequal axes $a$ (the tidal axis, oriented along a line from the center of the satellite to the center of the primary planet), $b$ (oriented tangent to the satellite’s orbit), and $c$ (the spin axis of the satellite). For an equilibrium satellite shape in synchronous orbit, the ratio $(a-c)/(b-c) = 4$ (with a small correction if the satellite is spinning rapidly), while the magnitudes of $(a-c)$ and $(b-c)$ depend on the rotation rate and internal density distribution (Murray and Dermott, 1999). If the rotation rate or magnitude of the tide changes, then so too will the shape of the satellite, generating global-scale patterns of stress.

3.1.2 Rigidity and effective elastic thickness

Outer planet satellites tend to have low surface temperatures, and as a result the near-surface region, hereafter referred to as the lithosphere, will be cold and rigid. As discussed above (Section 2.6), it is a convenient simplification to describe this near-surface region as a fully elastic layer overlying an inviscid interior. The thickness of this layer, $T_e$, is termed the effective elastic thickness of the lithosphere, and depends on the strength envelope (e.g., Figure 7.2) and thus the thermal gradient within the lithosphere. The lithosphere is able to support loads of some spatial scale over geological timescales, and is also the region in which permanent tectonic deformation (such as faulting) is recorded. The elastic thickness of the lithosphere $T_e$ controls the characteristic bending (or flexural) wavelength $\alpha$ at which deformation occurs in response to loading, and is given by (e.g., Turcotte and Schubert, 2002)

$$\alpha^4 = \frac{1}{3} \frac{ET_e^3}{(1 - v^2) \rho g},$$

(7.2)
where $E$ is the Young’s modulus, $\nu$ is Poisson’s ratio, $\rho$ the density of subsurface material being displaced (ductile ice or water), and $g$ is gravity.

Many of the expressions given below assume, usually implicitly, that the satellite’s figure is well described by hydrostatic equilibrium. This approximation is generally a good one for the large satellites. However, a thick lithosphere (either near-surface, or in a rocky core, if present) may invalidate the hydrostatic assumption, as is relevant to Ganymede (where non-hydrostatic mass anomalies have been inferred within the ice shell or core; Palguta et al., 2006) and potentially Callisto (McKinnon, 1997), and particularly for the mid-sized icy satellites of Saturn and Uranus (e.g., Rhea; Mackenzie et al., 2008). Long-wavelength loads are more readily supported on smaller planetary bodies due to the influence of membrane stresses (Turcotte et al., 1981). The most serious consequence of this result is that first-order interior structure models making use of the hydrostatic assumption may be incorrect (e.g., McKinnon, 1997). Increased rigidity will also reduce the amplitude of tidal and rotational deformation, as described below. A satellite with significant rigidity may also retain its shape from an epoch when the tidal and rotational characteristics were different from their present-day values (e.g., Iapetus), complicating present-day analysis.

### 3.1.3 Tides

With the significant exception of Hyperion at Saturn, all the regular satellites of the outer solar system are in (or very close to) synchronous rotation, showing the same face toward the primary planet. This is because tidal torques exerted by the primary on each satellite act to quickly tidally evolve any initially rapidly spinning satellites into this configuration (Peale, 1977; Murray and Dermott, 1999). Tides not only influence the orbital and rotational evolution of the satellites, but are a major source of stress and heat (e.g., Peale, 2003). Thus, one key manner in which satellite tectonics and geophysics differ from those of the terrestrial planets is in the influence of tides.

A satellite in orbit around its primary will experience a tidal bulge due to the difference in gravitational attraction of the primary from the near to the far side of the satellite. If a synchronously rotating satellite’s orbital eccentricity is zero, the tidal bulge will then be at a fixed geographical point, and of constant amplitude. The maximum amplitude $H$ of this static (or permanent) tidal bulge is given by (e.g., Murray and Dermott, 1999)

$$H = h_2 R_s \left( \frac{M_p}{M_s} \right) \left( \frac{R_s}{a_s} \right)^3,$$

where $R_s$ is the satellite radius, $a_s$ is the semimajor axis, $M_p$ and $M_s$ are the masses of the primary and satellite, respectively, and $h_2$ is a Love number that describes the
radial response of the satellite to a gravitational potential. The Love number $h_2$ has a value of $5/2$ for an idealized incompressible fluid body with uniform density $\rho$, but this is reduced as the rigidity (or shear modulus) $\mu$ of the satellite increases. For a homogeneous satellite, the reduction is a factor of $1 + \frac{10}{7} \tilde{\mu}$, where $\tilde{\mu} = \mu / \rho g R_s$ is a dimensionless measure of the importance of the rigidity (Murray and Dermott, 1999). A thin ice shell decoupled from the underlying material by an ocean has a low global rigidity, so a satellite with a thin floating ice shell can have a tidal response that approaches that of a fluid, provided that the entire interior can respond as a fluid on a sufficiently short timescale (e.g., Moore and Schubert, 2003).

For a synchronous satellite with non-zero orbital eccentricity, the magnitude of the tide varies as the satellite moves closer and farther from the primary planet, generating a time-varying radial tide. Moreover, the position of the sub-planet point will oscillate in longitude over the course of an orbit, and the amplitude and longitudinal position of the tidal bulge will thus vary in time, generating a librational tide. The combined result of the radial and librational tides is a time-varying diurnal tide with an amplitude given by $3eH$, where $e$ is the eccentricity (e.g., Greenberg et al., 2002). Table 7.1 gives nominal permanent and diurnal tidal bulge amplitudes for the satellites of the outer solar system. If the obliquity (the angle between the spin pole and the normal vector to the satellite’s orbital plane) of a satellite is non-zero, then the sub-planet point will librate in latitude as well as longitude. Such obliquity librations are potentially an additional source of tidal stress (Bills, 2005).

A time-varying diurnal tide leads to time-varying stresses within the satellite. The lithosphere can flex elastically from the diurnal tide, but if the viscoelastic interior has a Maxwell time (Section 2.5) comparable to the orbital period, then the tidal energy can be dissipated as heat within the satellite (e.g., Peale, 2003). Although the diurnal tidal strains are generally small ($10^{-5}$ or less), the periods are short, and thus the rate of tidal heating can be considerable, as at Io (Peale et al., 1979). Possible additional sources of tidal dissipation include obliquity variation (Bills, 2005), forced libration (Wisdom, 2004), or a large impact that triggers transient chaotic rotation (Marcialis and Greenberg, 1987).

Because the orbital speed of a satellite in an eccentric orbit varies while its rotation rate stays constant, the diurnal tidal component will both lead and lag the sub-planet point over the course of one orbit, generating a net non-zero torque (Goldreich, 1966; Greenberg and Weidenschilling, 1984). If this torque is not opposed by a sufficiently large permanent mass asymmetry, it can produce a rotation rate slightly faster than synchronous, so that the geographic location of the sub-planet point very slowly changes over time. As shown below, stresses due to both diurnal tides and nonsynchronous rotation can have important tectonic effects.
Table 7.1. Properties of the outer planet satellites

<table>
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<th>$R$ (km)</th>
<th>$M_s$ ($10^{20}$ kg)</th>
<th>$\rho$ (kg m$^{-3}$)</th>
<th>$a_s$ (10$^3$ km)</th>
<th>$P$ (days)</th>
<th>$e$</th>
<th>$H$ (m)</th>
<th>$3eH$ (m)</th>
<th>Tidal (GW)</th>
<th>Rad (GW)</th>
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Data from Yoder (1995) except for Saturn satellite radii and densities, which are from Thomas et al. (2007) and Triton and Charon from Perron et al. (2006). $H$ is the permanent tidal bulge assuming $h_2 = 2.5$. $3eH$ is the approximate magnitude of the diurnal tidal bulge; note that this is likely an overestimate for most satellites because the rigidity of their ice shells or cores will reduce $h_2$. “Tidal” is the tidal heat production assuming a homogeneous body with the Love number $k_2 = 1.5$ (again, a likely overestimate for small satellites) and dissipation quality factor $Q = 100$. “Rad” is the present-day radiogenic heat production assuming a chondritic rate of $3.5 \times 10^{-12}$ W kg$^{-1}$ for the rocky portions of the bodies. “Stress” is the approximate diurnal tidal stress given by $EeH/R$ where $E$ is Young’s modulus (assumed 9 GPa); see Section 3.2.1. The actual tidal stress varies according to location on the surface and position in the orbit. Asterisks denote bodies that are significantly non-spherical, and $R$ denotes retrograde rotation.
Figure 7.3. The time-averaged strain rate as a function of location for diurnal tidal flexing, normalized to a value of unity at the poles, modified from Ojakangas and Stevenson (1989a). Note that the strain rate is highest at the poles and lowest on the tidal axis.

3.2 Global stress mechanisms

Most of the stresses discussed in this section arise from changes in the satellite’s figure (subsection 3.1.1). Matsuyama and Nimmo (2008) give detailed expressions for the various kinds of stress-raising mechanisms enumerated here.

3.2.1 Diurnal tides

Subsection 3.1.3 discussed the important role of tides for a satellite in an elliptical orbit, as a function of its rigidity and semimajor axis. The maximum elastic stress generated by the diurnal tide is $\sim 3EeH/R_s$, where $E$ is the Young’s modulus of ice, and $H$ is the static tidal bulge amplitude (Equation 7.2); the diurnal tidal stresses are thus a strong function of distance from the primary planet and the total rigidity of the satellite (Table 7.1). Analytical expressions for the different components of the diurnal tidal stress tensor are complicated (Tobie et al., 2005), even in the case of a thin shell (Ojakangas and Stevenson, 1989a). Figure 7.3 shows a map of the average strain rate for a thin shell, demonstrating that strain rates and stresses are, on average, largest at the poles, and minimized at the sub- and anti-primary points. Note that the pattern is quite different for a satellite in which the shell is not thin (Segatz et al., 1988).

Diurnal stresses change with orbital position (Greenberg et al., 1998; Hoppa et al., 1999a). On the equator, the magnitudes and signs of the horizontal principal stresses change as the satellite progresses through an orbit, alternating between
compression and tension. Off the equator, the principal stress magnitudes and signs change, and the principal stress orientations also rotate through a diurnal cycle, counterclockwise in the northern hemisphere and clockwise in the southern. Consideration of the resultant out-of-phase variation in normal and shear stresses implies that contraction, extension, and (at high latitudes) opposite-sign horizontal principal stresses that promote strike-slip motions are all possible as a satellite orbits (Hoppa et al., 1999a). This motion can potentially lead to strike-slip displacement along faults, with a predicted preferred sense of left-lateral motion in the northern hemisphere and right-lateral motion in the southern hemisphere (Hoppa et al., 1999a), with the tendency for strike-slip motion increasing with higher latitude. For strike-slip displacement to occur, the fault friction $\mu_f$, overburden stress ($\rho g z$), and tidal normal stress ($\sigma_n$) must be sufficiently low so that (according to the Coulomb criterion),

$$|\tau_s| > \mu_f (\rho g z + \sigma_n),$$

(7.4)

where $\tau_s$ is the tidal shear stress acting on the fault, and $z$ is depth (e.g., Smith-Konter and Pappalardo, 2008).

### 3.2.2 Nonsynchronous rotation

As noted above, synchronously rotating satellites in elliptical orbits experience torques that may lead to slow nonsynchronous rotation (NSR) unless permanent mass asymmetries exist that resist the applied torques (Greenberg and Weidenschilling, 1984). In the case of floating ice shells maintained by tidal dissipation, there may be no equilibrium shell configuration that satisfies both thermal and rotational constraints, a situation that also can induce nonsynchronous rotation (Ojakangas and Stevenson, 1989a). Because satellites in synchronous rotation experience more impacts on the leading hemisphere than the trailing hemisphere, NSR might be detected by examining the longitudinal distribution of impact features (e.g., Zahnle et al., 2001).

If NSR occurs, the reorientation of the surface with respect to the tidal axis leads to surface stresses (Helfenstein and Parmentier, 1985; Leith and McKinnon, 1996; Greenberg et al., 1998). The principal horizontal stresses are both tensile in regions spanning $90^\circ$ of longitude and reaching to $\sim \pm 40^\circ$ latitude, centered $45^\circ$ westward of the tidal axis, as these regions are being stretched as they rotate up onto the tidal bulge. Similarly, the principal horizontal stresses are both compressive in similar zones centered $45^\circ$ eastward of the tidal axis, as these regions are rotating off of the tidal axis. Principal stresses with opposite signs, consistent with strike-slip motion, exist elsewhere across the satellite (Figure 7.4a). Stresses from NSR have been modeled as increasing with the angular amount of rotation, with the maximum
Figure 7.4. Horizontal principal stresses on the surface of a satellite arising from changes in the satellite’s figure. The +/+ symbol denotes areas where both principal stresses are compressive (which could lead to thrust faulting), −/− denotes areas where both principal stresses are tensile (which could lead to normal faulting), and +/− denotes areas with one compressive and one tensile principal stress (which could lead to strike-slip faulting). The magnitude of the stress, and thus whether faults are actually formed, depends on the magnitude of change. (a) Instantaneous nonsynchronous rotation of the ice shell to the east with respect to the tidal axis (at 0° and 180° longitude). (b) Polar wander of the ice shell by 90° with respect to the original rotation axis (solid circle). (c) Despinning. (d) Orbital recession of a tidally locked satellite.

tensile stresses described by

\[ \sigma = 6f \mu \left( \frac{1 + \nu}{5 + \nu} \right) \sin \theta = 3f E \frac{\sin \theta}{5 + \nu}, \]  

(Leith and McKinnon, 1996) where \( \mu \) is the shear modulus, \( \theta \) is the reorientation angle, and \( f \) is the satellite flattening in the tidal direction. For a hydrostatic body, this flattening is given by

\[ f = \frac{3R^3\omega^2}{2GMs}h_2, \] 

(7.5)
Planetary Tectonics (Leith and McKinnon, 1996). The flattening will be reduced if the satellite has significant rigidity (see subsection 3.1.2). The stresses due to nonsynchronous rotation will exceed the diurnal tidal stresses for $\theta > e$, or for a longitudinal reorientation of more than a few tenths of a degree for typical eccentricities. However, for likely nonsynchronous rotation rates, the strain rates generated will be much smaller than the strain rates caused by diurnal tides (Ojakangas and Stevenson, 1989a; Nimmo et al., 2007a).

Wahr et al. (2009) consider the combined NSR and diurnal stresses in a viscoelastic ice shell based on the gravitational potential fields of a rotating satellite and its primary planet, especially as relevant to Europa. The NSR stress affecting the ice shell is parameterized by rotation period $P_{\text{NSR}}$ relative to the Maxwell time $\tau_M$ of the ice, $\Delta = P_{\text{NSR}}/2\pi \tau_M$. For an ice shell with a sufficiently short rotation period or high viscosity ($\Delta \ll 1$), the ice behaves elastically on the NSR period and NSR stresses are large, overwhelming diurnal stresses; however, for an ice shell with a sufficiently long rotation period or low viscosity ($\Delta \gg 1$), the ice behaves visously so NSR stresses relax away and are small, permitting diurnal stresses to dominate. When $\Delta \approx 1/e$ for a satellite with orbital eccentricity $e$, NSR and diurnal stresses are of comparable magnitude, combining to influence the satellite’s tectonics.

### 3.2.3 Polar wander

In certain circumstances, the surface of a satellite may reorient relative to its axis of rotation (Willemann, 1984; Matsuyama and Nimmo, 2007). This process of polar wander is conceptually very similar to NSR, since it also involves motion of the surface relative to the satellite’s axes (Melosh, 1980a; Leith and McKinnon, 1996), but in addition to potentially moving through the tidal axis, the surface also moves through the polar axis, changing the stress pattern significantly. In general, the reorientation direction occurs roughly perpendicular to the tidal axis, because such paths are energetically favored (Matsuyama and Nimmo, 2007).

Floating ice shells maintained by tidal dissipation may undergo 90° reorientations about the fixed tidal axis due to increased ice thickness at the poles (Ojakangas and Stevenson, 1989b). Long-wavelength density anomalies may lead to smaller amounts of reorientation (Janes and Melosh, 1988; Nimmo and Pappalardo, 2006), as may volatile redistribution (Rubincam, 2003). Finally, large impacts may cause reorientation directly (e.g., Chapman and McKinnon, 1986) or due to the mass asymmetry from the creation of a new impact basin (e.g., Melosh, 1975; Murchie and Head, 1986; Nimmo and Matsuyama, 2007).

Polar wander induces tensile principal stresses (encouraging tension fractures or normal faulting) in the quadrant leading the reorientation direction because the region originally located along the spin axis must lengthen, and compressional
stresses (encouraging thrust faulting or folding) in the trailing quadrant because that region moves toward the shorter spin axis (Leith and McKinnon, 1996; Figure 7.4b). The maximum tensile stress that develops is given by Equation (7.5), with the value of the flattening $f$ depending on whether any reorientation of the tidal axis has occurred. If no such reorientation has occurred, the value of $f$ is simply one-third of the value given in Equation (7.6); in the more general case, the stresses will be larger and may be obtained from expressions given in Matsuyama and Nimmo (2008).

3.2.4 Despinning

Satellites that are initially rotating at a rate faster than synchronous will reduce their rotation rate, or despin, to that of synchronous rotation in timescales generally very short compared to the age of the solar system (e.g., Murray and Dermott, 1999). A reduction in spin rate means that the satellite will be less rotationally flattened; the resulting shape change gives rise to stresses – compressive at the equator as the equatorial bulge collapses and tensile at the poles as the poles elongate (Figure 7.4c). The maximum differential stress caused by despinning for a thin elastic shell in hydrostatic equilibrium is given by

$$
\sigma = 2 \left( \omega_1^2 - \omega_2^2 \right) \frac{R_s^3}{GM_s \mu} \left( \frac{1 + \nu}{5 + \nu} \right) h_2,
$$

(Melosh, 1977) where $\omega_1$ and $\omega_2$ are the initial and final angular rotation velocities. Note that if the satellite’s rigidity is large enough to reduce the rotational flattening (an effect included in $h_2$), the stresses will be reduced by a factor of $1 + \frac{19}{2} \tilde{\mu}$ for a homogeneous satellite (see Melosh, 1977; Squyres and Croft, 1986). Note also that spin-up is possible in principle in some circumstances (e.g., if the satellite undergoes differentiation, and if spin-up overcomes the strong despinning torque noted above); in this case the signs of all the stresses would be reversed.

3.2.5 Orbital recession and decay

Tidal dissipation within the primary planet causes satellites in prograde orbits outside synchronous altitude to recede, or spiral outward (as the Earth’s Moon is doing), and satellites in retrograde orbits or orbits under synchronous altitude to decay, or spiral inward. This will change the amplitude of the tidal bulge, and if the satellite is rotating synchronously, it will also change the spin rate. Most major satellites (with the notable exception of Triton) are in prograde orbits, and so generally undergo tidal recession. Recession causes a reduction in rotation rate, lengthening the $c$ axis, and a decrease in the static tide, shortening the $a$ axis. This combination of despinning and tidal bulge reduction causes a stress pattern in which a region around the sub-planet point experiences compressive stress, mid-latitudes experience horizontal shear stress, and the poles undergo tension (Figure 7.4d;
The maximum principal stress difference is twice the maximum stress given by Equation (7.7) (Melosh, 1980b). A satellite undergoing orbital decay experiences exactly the opposite changes in static tide and spin rate, and so experiences the same pattern of stress, but with the signs reversed.

Internal differentiation of a satellite may or may not cause significant global volume change (discussed in the next section), but since it concentrates mass toward the center of the body, it decreases the moment of interia. This will reduce $h_2$, and thus reduce both the tidal and rotational distortions. As long as the satellite remains in synchronous rotation, the surface stresses are the same as for orbital recession (Collins et al., 1999).

### 3.2.6 Volume change

A large number of different mechanisms can lead to volume changes within a satellite, and thus extensional or contractional features on the surface (Squyres and Croft, 1986; Kirk and Stevenson, 1987; Mueller and McKinnon, 1988). Such volume changes will generate isotropic stress fields on the surface. In many cases, the most important effect is that ice at high pressures (roughly $>0.2$ GPa) is considerably more dense than ice at low pressures. Thus when a satellite undergoes internal differentiation, high pressure ice in the interior is displaced by silicates (which are less compressible), leading to an overall increase in volume. The concomitant increase in surface area can reach several percent in the case of Ganymede and Callisto (Squyres, 1980; Mueller and McKinnon, 1988). Smaller amounts of expansion will occur as the satellite warms (e.g., through radioactive decay or tidal heating) and ice transforms from the high- to the low-pressure phase (Ellsworth and Schubert, 1983). These effects will be reversed if the satellite cools. Warming may also lead to silicate dehydration (Finnerty et al., 1981; Squyres and Croft, 1986), which in turn leads to expansion, unless dehydration occurs at pressures in excess of $\sim2$ GPa (e.g., Dobson et al., 2002). In the latter case the situation is more complicated: dehydration itself results in volume contraction, but eventual migration of released water to the satellite’s outer layers may lead to overall expansion.

Another potential source of expansion or contraction is the large density contrast between ice I and water. As water freezes to ice I, large tensile surface stresses can result (Cassen et al., 1979; Nimmo, 2004b). This effect is most important for small icy satellites and satellites with thin water layers, since oceans freezing in icy satellites with very thick water layers also produce higher density ice phases at the bottom. The volume change of water to high-pressure ice phases is opposite to that due to the freezing of ice I, and the combination of the two reduces the net volume change (cf. Squyres, 1980; Showman et al., 1997).
Solidification of initially molten iron cores in silicate bodies can lead to substantial contraction, as in the case of Mercury (e.g., Melosh and McKinnon, 1988), but is likely to be a minor effect in the case of icy satellites (Ganymede and Titan are possible exceptions).

The isotropic surface stresses $\sigma$ resulting from global expansion or contraction (McKinnon, 1982; Melosh and McKinnon, 1988) are given by

$$\sigma = 2\mu \frac{(1 + \nu) \Delta R}{(1 - \nu) R_s} = \frac{E}{1 - \nu} \frac{\Delta R}{R_s},$$

where $\Delta R$ is the radial contraction or expansion. A fractional change in radius of 0.1% gives rise to stresses of order 10 MPa. A closely related stress mechanism, which appears to be confined to Io, is burial of the surface by more recently erupted material (subsection 4.1.3; Schenk and Bulmer, 1998). This radially inward motion results in isotropic compressive stresses within the buried layers, as described by Equation (7.8).

Similar expressions arise for thermal expansion or contraction, where in this case the radial change is the product of the globally averaged thermal expansivity and the temperature change (Ellsworth and Schubert, 1983; Zuber and Parmentier, 1984; Hillier and Squyres, 1991; Showman et al., 1997). A global temperature change of 100 K gives rise to stresses of order 30 MPa for a volume thermal expansivity ($\alpha_V$) of $1 \times 10^{-4}$ K$^{-1}$. Thickening ice shells result in stresses due to both volumetric and thermal effects, but the former dominate (Nimmo, 2004b). Direct thermoelastic stresses will arise in cooling or warming lithospheres with or without underlying global volume change, since the lithospheres are laterally confined (Turcotte, 1983; Hillier and Squyres, 1991). Maximum stresses are $E\alpha_V \Delta T/3 (1 - \nu) \sim 3 \times (\Delta T/10$ K) MPa for water ice near 150 K, but decline for colder ice since $\alpha_V$ is strongly temperature dependent (Petrenko and Whitworth, 1999).

### 3.3 Local stress mechanisms

Satellites are quite likely to have non-axisymmetric structures, in which case some of the above global mechanisms may lead to local deformation. For instance, local zones of weakness can result in enhanced tidal dissipation and deformation (e.g., Sotin et al., 2002), as may be occurring at the south pole of Enceladus. However, there are also stress-generating mechanisms that are intrinsically local in character, which will be itemized here.

#### 3.3.1 Convection

Thermal convection is a potentially important process on icy satellites (e.g., McKinnon, 1999), since it will substantially alter the thermal evolution of such
bodies, including the potential for subsurface oceans, and may also cause local
deformation (e.g., Nimmo and Manga, 2002). For large icy satellites, layers of
high- and low-pressure ice may convect separately (e.g., McKinnon, 1998).
Whether convection occurs depends on the ice shell thickness, local gravity, and
most importantly, the viscosity of the ice. As outlined in Section 2.4, the viscous-
ity is both temperature- and grain-size dependent, and the biggest uncertainty in
assessing whether convection occurs is due to uncertainties in the grain size (Barr
and Pappalardo, 2005). The characteristic stresses generated by convection on
the overlying lithosphere depend on the temperature difference available to drive
convection; for strongly temperature-dependent viscosity (Solomatov and Moresi,
2000) the stresses are given by
\[ \sigma \approx 0.03 \alpha_\rho \rho g \Delta T_{rh} \delta_{rh}, \]  
(7.9)
where \( \Delta T_{rh} \) and \( \delta_{rh} \) are the rheologically controlled temperature drop and boundary
layer thickness. The latter controls the length scale of convective features, and
depends on the vigor of convection; for ice it is typically of order a few km
(see Solomatov and Moresi, 2000). The temperature contrast \( \Delta T_{rh} \) for Newtonian
rheologies is given by
\[ \Delta T_{rh} \approx 2.4 R T_i^2 \frac{Q}{Q}, \]  
(7.10)
where \( R \) is the gas constant, \( T_i \) is the interior temperature, and \( Q \) is the activation
energy. Assuming \( Q = 60 \text{ kJ/mol} \) and \( T_i = 250 \text{ K} \), we have \( \Delta T_{rh} \approx 20 \text{ K} \). The
resulting stresses, from Equation (7.9), are of order 1 kPa for a boundary layer
thickness of 10 km and a gravity of 1 m s\(^{-2}\). These stresses are generally small
compared to the stresses arising due to other mechanisms. Furthermore, even
though larger stresses than given by Equation (7.9) are generated near the surface
in a thick, stagnant lid (Solomatov, 1995; Solomatov and Moresi, 2000), the lid is
cold and immobile, which further reduces surface deformation (e.g., Nimmo and
Manga, 2002). In general, therefore, one might not expect convection in an ice
shell to generate identifiable surface features.

Convective stresses on silicate bodies tend to be larger, because the higher
operating temperatures overwhelm the effect of the larger silicate activation energy
in Equation (7.10), and the rheological length scales are typically greater. The
kilometer high long-wavelength convective uplifts of the kind seen on Earth or
Venus are typical of silicate convection, but not of that in ice shells.

Under certain circumstances, convection can lead to larger stresses. For instance,
satellite differentiation may lead to a deep ice layer with a high potential temper-
ature, which drives large-scale convective overturn and generates large tensile
stresses (Kirk and Stevenson, 1987). If the near-surface ice is sufficiently weak
to undergo yielding, then larger surface deformations and stresses may result (Showman and Han, 2005; Han and Showman, 2008; Barr, 2008). Compositional convection may also give rise to large stresses, because density contrasts driven by compositional variations can be much larger than those driven by thermal differences. It is possible that thermal convection, or other local heating mechanisms, can generate compositional contrasts by preferentially melting salt-rich, dense phases (Nimmo et al., 2003a; Pappalardo and Barr, 2004). However, unlike thermal convection, composition-driven overturn will only happen once, unless there is some mechanism continuously generating new compositional contrasts.

3.3.2 Lateral pressure gradients
For a floating ice shell, if isostatically compensated lateral shell thickness variations exist, then lateral pressure gradients occur. The force per unit length (e.g., Buck, 1991) is given by

\[ F \approx g \Delta \rho t_c \Delta t_c, \]

where \( \Delta \rho \) is the density contrast between shell and underlying ocean and \( t_c \) and \( \Delta t_c \) are the mean shell thickness and the thickness variation, respectively. If these forces are distributed uniformly across the entire crust, the stress is simply \( F/t_c \). To produce a 0.1 km variation in elevation requires a \( \Delta t_c \) of about 1 km if \( \Delta \rho = 100 \text{ kg m}^{-3} \). The resulting stress distributed across the entire ice shell is 10 kPa for \( g = 1 \text{ ms}^{-2} \).

One consequence of these pressure gradients for a floating ice shell is that lateral flow of the low-viscosity ice at the base of the shell will occur, removing shell thickness contrasts (Ojakangas and Stevenson, 1989a; Nimmo, 2004c). The resulting vertical motions of the ice shell may result in the generation of surface stresses. Lateral flow in the shell is also important in determining how ice shells respond to tensile stresses, and in particular whether wide or narrow rifts develop (Nimmo, 2004d).

3.3.3 Flexure
If an ice shell has long-term rigidity, then loads emplaced on it will be partially supported by the strength of the lithosphere. The deflection of the lithosphere in response to the load will in turn result in stresses. For a lithosphere with an effective elastic thickness \( T_e \), when the load wavelength \( \lambda \) is much less than the satellite radius, the maximum deflection of the ice shell \( w_0 \) in response to a sinusoidal load that generates a final topography \( h_0 \) is given by (Turcotte and Schubert, 2002)

\[ w_0 = h_0 \frac{\rho}{\Delta \rho} \left( 1 + \frac{k^4 E T_e^3}{12(1 - \nu^2)g \Delta \rho} \right)^{-1}, \]
where $\Delta \rho$ is the density contrast between the shell and the water beneath and $k$ is the wavenumber ($= 2\pi / \lambda$). For long-wavelength loads, or if the elastic thickness is zero, then this result simplifies to the usual isostatic case. Larger elastic thicknesses result in reduced deflections.

For this same load, the maximum stresses experienced are given by (Turcotte and Schubert, 2002)

$$\sigma = \frac{E}{(1-v^2)} \frac{T_e}{2} k^2 w_0.$$  \hfill (7.13)

In the simplified elastic model, bending stresses at any location relative to the point of greatest deflection are maximized at the surface and base of the elastic layer, respectively, but with opposite signs.

### 3.3.4 Impacts

Impacts obviously generate transient, large local stresses. Sufficiently large impacts may induce lateral flow, generating tectonic features significantly outside the original impact area (McKinnon and Melosh, 1980). In some cases, focusing of impact-generated waves may in principle also result in tectonic features being generated at the impact antipode (Bruesch and Asphaug, 2004; Moore et al., 2004a). Over the longer term, impact sites may act as zones of weakness that affect the spatial distribution of other tectonic features (cf. McKenzie et al., 1992). Impact basins may also generate stresses through secondary mechanisms such as polar wander (subsection 3.2.3) or lateral flow in the ice shell (subsection 3.3.2).

### 4 Io

Io is the only outer planet satellite with a rocky (i.e., water-ice free) surface, and it is also the most geologically active of the satellites. Io and the Earth are rivals for the title of most geologically active body in the solar system, but the two bodies could not be more different in how they exhibit their dynamic personalities. It has become the paradigm that planets release excess internal heat through volcanism and tectonism. On Earth, the main mechanism of heat loss is by the creation of new lithospheric plates at mid-ocean ridges. These oceanic plates cool conductively and thicken as they move laterally toward subduction zones, leading to the horizontal and vertical conveyor belt of crustal recycling known as plate tectonics.

Io is very similar to the Earth’s moon in size and density, and is inferred to have a dominantly silicate composition (with significant differences such as abundant sulfur). Thus, when Voyager discovered Io’s intense global volcanism in 1979 (Peale et al., 1979; Smith et al., 1979), the lack of long linear mountain chains was also immediately obvious and a source of some consternation. How could such an
active volcanic world lack the organized tectonic network observed on Earth? If Io’s interior is so hot and active, why is there no plate tectonics? It seems that Io has its own unique form of vertical crustal recycling.

4.1 Tectonic features on Io

Io may be dominated by volcanism, but it certainly has tectonic features. The most common tectonic features are volcanic calderas (e.g., Radebaugh et al., 2001), referred to as paterae on Io. Calderas, fault-bounded quasi-circular depressions, form as collapse features over magma chambers and are usually associated with extensive lava flows. There are over 500 volcanic centers on Io, many of which are associated with calderas of some type (Figure 7.5). In addition, there are a few curvilinear fractures on Io, most likely volcanic fissures, flanked by short lava flows a few tens of kilometers long. These lava flows form symmetric butterfly-like patterns on the surface, again indicating fissure eruptions of lava. Calderas and fissures are directly associated with volcanism and are relatively well understood in that context. In the rest of this section we will focus on tectonic features that are not obviously of direct volcanic origin.

4.1.1 Fractures

Narrow curvilinear features are scattered across Io’s volcanic plains. These features are dwarfed by more prominent neighboring volcanoes and mountains, and so they
have been largely ignored in the literature. Typically 1–3 kilometers across, some of these features can stretch for over 500 kilometers in length (Figure 7.6). Their morphology is consistent with simple extensional fractures or graben.

The formation of these fractures may be related to global tidal stresses. Io differs from most other worlds in the intense daily tidal deformation of its surface. Due to its close proximity to giant Jupiter and its orbital resonance with neighboring moons Europa and Ganymede, Io’s orbit is eccentric and its solid surface experiences daily tides of up to $\sim 0.1$ km (Table 7.1), leading to repetitive surface strains of $10^{-4}$ or greater. These tides flex and stress the lithosphere and can cause it to fracture (as also occurs extensively on neighboring Europa – see Section 5.1). However, no correlation has yet been found between fracture orientation and the predicted stresses resulting from tides. The situation can be confused if the features formed at different times or if the stress pattern shifts due to nonsynchronous rotation of the lithosphere (Milazzo et al., 2001).

Alternatively, curvilinear or concentric extensional fractures could be related to local loading of planetary lithospheres. On Io, this could be the result of construction of volcanic edifices or global convection patterns forming localized sites of upwelling and downwelling (e.g., Tackley et al., 2001). However, constructional volcanic edifices are quite rare on Io (Schenk et al., 2004a) and convective stresses on Io are likely to be quite small (Kirchoff and McKinnon, 2009). On Io, constant global resurfacing by lavas and deposition from volcanic plumes can locally

Figure 7.6. Examples of graben-like features on the surface of Io (marked by arrows). These features are typically a few kilometers wide. None of these features were observed at resolution better than a few hundred meters per pixel. Although they are likely to be extensional in origin, their relationship to local or global stress fields is unknown. Note how the graben-like features cross the cuspate scarp of an elevated plain (either an erosional scarp or a lava flow front). Illumination is from the right, and the dark stripe is a data gap.
erase tectonic patterns of this sort, in part or entirely. In the future, more sophisticated mapping and analysis will be required to explain these fractures, including correlations with topographic and gravity information.

4.1.2 Ridges

Sets of small-scale ridges are observed in certain high-resolution Galileo images of Io. They are short, typically a few kilometers in length, closely spaced (∼1 km apart), and of modest amplitude (probably less than 100 m in height). What is remarkable is their consistent orientation within each set across several tens of kilometers in extent. Bart et al. (2004) find that ridge orientations are not inconsistent with dominant stress directions during the diurnal tidal cycle, but these orientations vary during the cycle, especially for locations away from the equator, and ridge directions are not uniformly consistent with either a compressional or tensional tidal stress origin. Bart et al. (2004) list specific mechanisms that might account for the ridges: compressional folding, infill of open tension fractures, and differential compaction of a porous surface layer. Whatever the mechanism, it is clear that these features are young even by Io standards, since they are unburied, and they involve deformation of very shallow units or layers. Understanding their global distribution and their relation to other features on Io will require better and more consistent image coverage.

4.1.3 Mountains

The most prominent tectonic features on Io are the ∼150 mountains scattered across the surface (e.g., Schenk et al., 2001a; Turtle et al., 2001). These peaks average around 6 kilometers high, but the highest, Boosaule Montes, towers 17 kilometers above the flat volcanic plains surrounding it (Schenk et al., 2001a). Most mountains on Io are a few hundred kilometers across or less. There are several curious aspects to these mountains. Much like the numerous volcanic centers, each mountain keeps a respectable distance from its neighbors. Mountains rarely line up in close chains or clusters. Mass wasting of mountain faces is also evident. Several cases have been identified whereby large portions of mountains have failed by landslide, including one of the largest known landslides in the solar system (e.g., Schenk and Bulmer, 1998). More frequently, evidence of downslope creep of material is observed, as mountains apparently are slowly deformed by gravitational collapse (Turtle et al., 2001). Some mountains are merely isolated promontories only a few kilometers across. Given Io’s prodigious volcanic outpourings, it would not be surprising if we are observing mountains in various stages of burial.

With the exception of four known and easily recognizable conical shield volcanoes (Moore et al., 1986; Schenk et al., 2004a), no evidence has been found for
volcanic flow or caldera formation on the slopes or tops of any Ionian mountains. Thus, Io’s mountains must form by tectonic deformation and not by volcanic construction. This is consistent with the variety of mountain morphologies observed (Schenk et al., 2001a; Turtle et al., 2001), including flat-topped mesas, small narrow peaks, craggy massifs, rounded “eroded” massifs, elongate ridges, and asymmetric “flatirons” (Figure 7.7). These last morphologies are consistent with fault formation and are a key to understanding the origin of mountains on Io.

Flatiron mountains on Earth, such as those along the front range of the Rocky Mountains in Colorado, are associated with the tilting of thick, erosion-resistant layers by some combination of folding and thrust faulting, when two large masses of crust are forced together or upward by horizontal compression. Such flatirons on the Earth are fluvially dissected into flat triangular pieces, but on Io they appear
as solid flat walls tilting up from the surrounding plains. It is difficult to envision how extension would uplift lithospheric blocks 10 or more kilometers above the surrounding plains, so the apparent presence of large tilted blocks of crust on Io suggests that compression may indeed be driving the formation of Io’s mountains (Schenk and Bulmer, 1998).

Understanding the source of a global compressive stress field requires an understanding of how Io’s global volcanism is processed. Io’s volcanism is so intense that an estimated >1 mm to ~1 cm global layer of new lava is deposited on the surface each year (Johnson and Soderblom, 1982; McEwen et al., 2004). Volcanic eruptions are local phenomena, so this estimate is a global average calculated over millions of years, but the global effect of this volcanism is that the entire surface is continually being renewed over millennia (hence the lack of impact craters). It is clear from global mapping (e.g., Crown et al., 1992) that Io does not have lateral tectonics like the Earth: there are no subduction zones or spreading ridges. As a result, these new volcanic deposits must go somewhere, and on Io the only place to go is down. The new deposits force the older cooled lavas downward into the interior, and the result is a remarkably cool lithosphere of potentially great thickness (>25 km), as the recycling time of the crust (thickness divided by burial rate) is faster than the thermal equilibration time of the crust, for all but very thin crusts (O’Reilly and Davies, 1981; McKinnon et al., 2001). It is this rigid lithosphere that supports the large surface loads implied by the mountains (subsection 3.1.2).

The cool lithosphere was recognized shortly after the Voyager discoveries. A more subtle and elegant consequence of Io’s regime of “heat piping” and vertical recycling became clear much later. The volcanic layers (i.e., units of similar age) that were once at the surface no longer fit the decreasing surface area as they descend into the interior, due to the simple geometric reality that the surface area of a sphere decreases as it shrinks in radius. The unavoidable result is that these older buried layers must be subject to increasing lateral compression as burial continues. This model can be thought of as an onion that grows from the outside rather than from the inside. Rocks under compression can fail by ductile shortening or by brittle failure. Lower portions of Io’s crust may sometimes break under the strain, thrusting a large intact block upward, locally relieving some of the stress and forming a mountain on the surface (Schenk and Bulmer, 1998). Thus, mountain building on Io can be thought of as a direct consequence of Io’s high global volcanic resurfacing rates.

A further corollary to the vertical recycling model described above suggests that fluctuations in Io’s volcanic resurfacing rate cause variations in the heat flow at the base of the lithosphere (McKinnon et al., 2001; Kirchoff and McKinnon, 2009). These heat flow “pulses” might then increase stresses in the lower crust, triggering
mountain formation. That is, the close proximity of a deep, cool lithosphere, whose temperature gradient is suppressed by downward advection, to a subjacent hot, tidally heated asthenosphere, is potentially thermally unstable. The compressive thermoelastic strains that could result, for a $\Delta T \sim 1000$ K, are of order 1% (subsection 3.2.6), comparable to those which result directly from crustal subsidence, $z/R$, when $z$ is tens of kilometers. In both cases, the resulting stresses easily exceed Byerlee’s rule in compression (Jaeger et al., 2003; Kirchoff and McKinnon, 2009). The unsolved parameter is the importance of the thermal stresses relative to the subsidence stresses that most assuredly occur on Io.

The lack of a global thrust fault network or pattern can also be understood in the vertical tectonics scenario. If the global subsidence rate is more or less uniform across the surface, as suggested by the apparently uniform distribution of volcanic centers, then compressive stresses will also be nearly uniform within the lithosphere (see the next section for exceptions). Uniform compressive stresses in a thick lithosphere would likely be relieved locally, chiefly through thrust faulting triggered at local discontinuities or other failure points such as volcanic intrusions, preexisting faults, compositional boundaries, or areas of high thermoelastic stress, and mountain formation will occur in globally distributed “random” locations over time. Moreover, both the subsidence stress produced by downward vertical crustal advection, and compressive thermal stresses at depth that would develop during a local slowdown of volcanic advection of heat, would generate substantial bending moments, which would be released by mountain formation. Kirchoff and McKinnon (2009) suggest that the spacing of Ionian mountains and their relative isolation from each other are flexurally controlled.

4.2 Global distribution of mountains and volcanoes

Mapping of the distribution of mountains and volcanic centers shows that they are not exactly uniform, and there is a subtle but significant global pattern. There are two large areas on the surface where mountains appear to be significantly more numerous than the global average (Schenk et al., 2001a). These areas are not sharply defined, but rather are broad zones of increased mountain formation on opposite sides of the globe from each other, displaced about 75° west of the current sub- and anti-Jovian regions. A similar antipodal pattern emerges for the global distribution of volcanic centers (as distinct from individual flows that may emanate from specific centers) (Schenk et al., 2001a; Radebaugh et al., 2001). However, the distribution of mountains and volcanic centers are significantly anticorrelated: the areas where mountains are less frequent are the areas where volcanic centers are more numerous (McKinnon et al., 2001; Kirchoff, 2006). This seems counterintuitive if we believe the subsidence tectonics model described above. Areas with more
volcanism should experience more subsidence and hence more mountain building. Io is not so simple, apparently, and several factors could explain this anomaly. If internal convection is active on Io (Tackley et al., 2001) in a globally symmetric fashion, stress fields over rising and descending mantle plumes could potentially modify the compressive stresses due to subsidence, by enhancing volcanism and suppressing mountain building. Similarly, if nonsynchronous rotation occurs on Io (as proposed for Europa, see subsections 3.2.2 and Section 5.1), then this would impose an additional globally symmetric stress field within Io and could inhibit (or encourage) mountain formation in antipodal areas. Higher rates of volcanism could lead to thicker crust in those areas, although this should lead to more instead of less mountain formation.

The stresses generated by high Rayleigh number convection, as in Tackley (2001), should be quite modest in comparison with subsidence stresses (Kirchoff and McKinnon, 2009) because internal viscosities are small, and are thus unlikely to influence mountain formation. On the other hand, nonsynchronous rotation stresses could reach levels as great as 220 MPa (Equation 7.5) for $E = 65$ MPa, and $h_2 = 2.3$ (Schubert et al., 2004). Such stresses would certainly be important for Io’s lithosphere (especially its upper portion), but a large amount of nonsynchronous rotation, approaching $90^\circ$, would be necessary to match the actual positions of mountain and volcano concentrations, which would preferentially form in compression and tension, respectively. Strong mountain and volcanic feature orientation patterns would also be predicted if nonsynchronous rotation was an important source of stress, but such patterns have not yet been reported.

An alternative view of the anticorrelated mountain and volcanic feature distribution patterns invokes not spatially variable stress sources, but temporally varying ones. For example, regions that experience a slowdown or cessation of volcanism, and thus a slowdown in vertical crustal advection, should experience a buildup of compressive thermal stresses in the lower crust and possibly crustal thinning (McKinnon et al., 2001; Kirchoff and McKinnon, 2009). This would lead to a period of enhanced mountain formation, which could last until the volcanic “heat pipe” network was able to reestablish itself (which could in turn depend on the creation or activation of mountain-forming faults). In the context of the presently observed degree-2 patterns of volcanism and mountain building, slow nonsynchronous rotation is a possible cause, in that the present regions of volcanic concentration are centered on the sub- and anti-Jovian points, which is where tidal heating should be maximized for an asthenosphere-bearing Io (e.g., Tackley, 2001). Regions $90^\circ$ away would be former regions of enhanced volcanism, now diminished, that are now undergoing a period of enhanced orogenesis.

Volcanism and mountain formation may be intimately related on a local scale as well. Several instances have been identified in which volcanic paterae occur
along the outer margins of mountains, in some cases cutting into the basal scarps (Figure 7.7). In a few instances, both ends of a mountain are abutted by volcanic centers (Turtle et al., 2001; Schenk et al., 2001a; Jaeger et al., 2003). Normally, volcanism should be difficult under the compressional lithospheric stress regime described above, as strong compression would inhibit dike propagation, but local faulting could relieve these stresses enough to permit magma ascent. On the other hand, localized heating from a magma body could form a lithospheric discontinuity which could trigger faulting and stress release. The physical link between mountains and the formation of eruptive centers is not well understood, however. There are numerous counterexamples of isolated mountains that have formed hundreds of kilometers from the nearest volcano, which is remarkable given the ubiquitous presence of volcanic centers on Io.

After initially confounding us with their mere presence, it is now apparent that Io’s isolated but prominent mountains are intimately linked to its high heat flow and volcanic resurfacing rates. Most of the mountains have probably formed due to global inward subsidence of the volcanic surface and resulting compressional stress in the lower crust. Following uplift, additional volcanism may occur along faults formed by the mountains, and erosive processes such as downslope creep, landslides, and volcanic burial will begin to degrade mountain landforms, and ultimately return them from whence they came.

5 Active icy satellites

In this section, we discuss three ice-covered satellites that stand out from the others, in terms of having very young surfaces and even displaying current activity on their surfaces.

5.1 Europa

Europa, the smallest of Jupiter’s Galilean satellites, is a dynamic body so covered with overlapping linear ridges that it has often been compared in appearance to a giant ball of twine. From analysis of data from the Voyager and Galileo spacecraft, a comprehensive picture is emerging of how Europa’s global stress mechanisms, interior structure, and surface geology are inherently linked. Tectonism provides this link, and therefore an understanding of Europa’s tectonics is crucial for understanding the great unresolved questions of where tidal heat is deposited in Europe and the thickness of the ice shell. In this section we will address the tectonics of specific Europan landform types (fractures, ridges, bands, folds, domes and lenticulae, chaos, and large impacts), the global tectonic patterns and their links to stress mechanisms (diurnal stressing and nonsynchronous rotation), and the question of
Tectonics of the outer planet satellites

Figure 7.8. Typical tectonic features on Jupiter’s satellite Europa. (A) denotes a trough with no flanking ridges, trending east–west; (B) denotes two large examples of the ubiquitous double ridges; (C) denotes a ridge complex. North is to the top, and the illumination is from the right.

active tectonism (see Greeley et al., 2004, for a more general discussion of Europa surface geology).

5.1.1 Tectonics of Europa’s landforms

5.1.1.1 Isolated troughs The simplest of Europa’s landforms are individual linear to curvilinear troughs (Figure 7.8a). They are generally \( \sim 100 – 300 \) m wide and V-shaped in cross section, and they can be several to hundreds of kilometers long. Their rims can be relatively level, or raised with respect to the surrounding terrain. Interactions between propagating troughs are consistent with linear elastic fracture mechanics models of perturbed stress fields around fracture tips (Kattenhorn and Marshall, 2006). All of these characteristics suggest an origin as tension fractures.

The widths of visible troughs imply that they have widened subsequent to the original fracturing. The surface width \( w \) and depth \( z \) of a fracture can be related through the Young’s modulus \( E \) as \( w \sim \rho g z^2 / E \) (Nur, 1982). For \( E \sim 10^9 \) Pa (reduced from the solid-state value; Nimmo et al., 2003b), crevasse depths of
100 m, 1 km, and 10 km could produce surface widths ∼1 cm, 1 m, and 100 m, respectively. A significant portion of the width of isolated troughs may be due to mass wasting of debris from the sides of the trough and/or due to tectonic movement along the trough, e.g., from upbowing or strike-slip movement (discussed below).

5.1.1.2 Normal faults Normal faults are ubiquitous in extensional regions on Earth (e.g., Jackson, 1989), but have proved difficult to detect on Europa owing to limited imaging and topographic coverage. Imbricate normal faults have been inferred within Europa’s bands (Figueredo and Greeley, 2000; Prockter et al., 2002; Kattenhorn, 2002), as discussed in subsection 5.1.1.4. Two prominent normal faults have been described by Nimmo and Schenk (2006), with vertical offsets of several hundred meters and flexural flanks implying elastic thicknesses in the range 0.15–1.2 km. The maximum displacement/length ratio is ∼0.02, comparable to values on silicate bodies (e.g., Watters et al., 2000). The driving stresses implied by the existence of these faults are several MPa, and the derived near-surface shear modulus is apparently lower than that of intact ice, perhaps as a result of porosity and/or fracturing (Section 2.2).

5.1.1.3 Ridges Ridges are Europa’s most ubiquitous landform (Figure 7.8), yet their origin remains poorly understood (Pappalardo et al., 1999; Greeley et al., 2000, 2004). They most commonly take the form of a double ridge, i.e., a ridge pair with a medial trough (Figure 7.8b). Understanding their mode of formation is important to the satellite’s geophysical state, including the presence and distribution of liquid water. Suggested classification schemes (Greenberg et al., 1998; Head et al., 1999; Figueredo and Greeley, 2000) indicate a morphological transition from isolated troughs, to double ridges, to wider ridge complexes that commonly show a series of subparallel component ridges (Figure 7.8c). This morphological progression suggests an evolutionary sequence in which isolated troughs evolve into double ridges, then into more complex ridge morphologies. Some double ridges instead transition into wider bands, apparently as they are pulled apart along their axes (see subsection 5.1.1.4).

Double ridges have average widths of a few hundred meters, with prominent ridges being ∼2 km wide. Double ridges are characterized by a continuous axial trough that is V-shaped in cross section, and not as deep as its flanking ridges are tall. Ridge slopes are near the angle of repose (Kadel et al., 1998). Preexisting topography is sometimes partially recognized up the outer flanks of ridges, with the most prominent topography extending to near the top of the outer flanks (Head et al., 1999). Mass wasting is prevalent along ridge flanks, with the debris apparently draping over preexisting terrain (Sullivan et al., 1999).

Ridge complexes are commonly ∼5–10 km wide, consisting of multiple subparallel lineations and ridges, which can interweave or merge along their trends
(e.g., Figueredo and Greeley, 2000). Ridge complexes may be transitional between narrower double ridges and wider bands (subsection 5.1.1.4).

Many double ridges and ridge complexes show evidence for strike-slip motion along them, a characteristic not shared by isolated troughs (Hoppa et al., 1999a). Reconstruction of preexisting lineaments provides evidence for extension across some ridges (Tufts et al., 2000) and minor contraction across others (Patterson et al., 2006; Bader and Kattenhorn, 2007).

Some ridges are flanked by topographic depressions and/or fine-scale fractures (Tufts et al., 2000; Billings and Kattenhorn, 2005; Hurford et al., 2005). The downwarping adjacent to the ridge suggests loading of the lithosphere either from above (due to the weight of the overburden) or from below (due to withdrawal of subsurface material). With the assumption that ridges have loaded the surface, Billings and Kattenhorn (2005) use the distance to ridge-flanking cracks to estimate an effective elastic thickness of \( \sim 0.2 \) to 3 km, and Hurford et al. (2005) fit photoclinometric profiles to derive an average elastic lithospheric thickness of \( \sim 0.2 \) km (results which are sensitive to the assumed value of \( E \)). Such thin elastic lithospheres near cracks and ridges may not be representative of Europa’s ice shell as a whole, however.

At low solar incidence angle, it is apparent that some ridges have diffuse dark material infilling the flanking topographic depressions. The dark flanks that define these “triple bands” may have been created by ballistic emplacement of dark material entrained in gas-driven cryovolcanic eruptions (e.g., Crawford and Stevenson, 1988); alternatively, the flanks may be thin dark lag deposits due to sublimation of surface frosts and local concentration of refractory materials adjacent to a subsurface heat source associated with the ridge formation (Fagents et al., 2000).

Several models have been proposed for the origin of Europa’s ridges (cf. Pappalardo et al., 1999; Greeley et al., 2004), and these models have various implications for the presence and distribution of liquid water at the time of ridge formation. Some models invoke a shallow subsurface ocean; some rely on the action of warm mobile ice with perhaps an ocean at depth; some imply that liquid water exists in the shallow subsurface on at least an intermittent basis.

**Tidal squeezing.** Greenberg et al. (1998) propose that fractures penetrate completely through Europa’s ice shell, and open and close in response to changing diurnal stress, allowing water and icy debris to be pumped toward the surface with each tidal cycle to build ridges. In this model, opposing lithospheric blocks pull apart along \( \sim 1 \) m wide cracks that penetrate through the entire ice shell to an ocean below. As they pull apart, water rises up into the cracks hydrostatically. Europa’s diurnal stress cycle soon reverses the strain direction and closes the crack again, driving material to the surface. This “pumping” mechanism every 3.5 days is envisioned to pile up enough ice and slush on the surface adjacent to the crack to form double ridges. Because this model explicitly assumes that the entire ice
shell of Europa is penetrated by cracks, it envisions a very thin ice shell, one which more readily allows cracks to penetrate completely through the ice (Golombek and Banerdt, 1990; Leith and McKinnon, 1996). While Crawford and Stevenson (1988) discuss the difficulty of cracking from the base of the ice shell upward through warm ductile ice at the base of the ice shell, Lee et al. (2005) suggest that fractures formed at the top of an ice lithosphere can more easily penetrate downward through the entire plate because stress is concentrated at their tips. However, a complication for the tidal squeezing model is that water in narrow cracks is expected to freeze faster than the tidal cycle (Nimmo and Gaidos, 2002).

**Linear volcanism.** Kadel et al. (1998) propose that double ridges are linear volcanic constructs, built of debris associated with gas-driven fissure eruptions. Volcanic models suggest that volatiles such as CO$_2$ or SO$_2$ are capable of driving eruptions, overcoming the negative buoyancy of water relative to ice (Fagents et al., 2000). Like the tidal pumping model, the volcanic model is challenged in presuming open conduits extend from a subsurface ocean to the surface. It is possible that shallow melt chambers feed conduits instead, or that volatiles have driven pinched off water-filled cracks toward the surface (Crawford and Stevenson, 1988). However, this model also has difficulty accounting for the great linearity and continuity of Europa’s ridges, as terrestrial volcanic ridges tend to pinch and swell, due to eruption and coalescence of material into discrete eruption centers.

**Dike intrusion.** Melosh and Turtle (2004) propose that ridges form by intrusion of melt water into a shallow vertical crack to build a double ridge. In this model, melt intrudes into the shallow subsurface within dikes and subsequently freezes, causing outward and upward plastic deformation of the near-surface to create a ridge.

**Compression.** Sullivan et al. (1998) propose that ridges are contractional structures, deformed along plate boundaries. Reconstruction of preexisting structures suggests that compression is a viable model for some ridges (Sarid et al., 2002; Patterson et al., 2006; Bader and Kattenhorn, 2007). Contractional strain at ridges can help to compensate for the large degree of extensional strain represented by Europa’s bands.

**Linear diapirism.** Head et al. (1999) propose that double ridges form in response to cracking and consequent diapirc rise of tabular walls of warm ice, which intrude and uplift the surface to form ridges. This model suggests that cracks penetrate down to a subsurface ductile ice layer, rather than through the ice shell. Warm subsurface ice moves buoyantly into the fracture, aided by tidal heating concentrated along the fracture (Stevenson, 1996; Gaidos and Nimmo, 2000; Nimmo and Gaidos, 2002). The process is envisioned as analogous to the rise of tabular “salt walls” that rise along extensional fractures on Earth (e.g., Jenyon, 1986) and can cause intrusive uplift. In some cases the trend of the preexisting topography appears to
be deflected at a ridge as the ridge flank is encountered, consistent with the idea that ridge flanks may have formed by upwarping of the preexisting surface rather than by volcanic construction. As with the caveat regarding the linear volcanism model, the symmetry, uniformity, continuity, and great lengths of ridges on Europa are unmatched by any linear diapir on the Earth.

**Shear heating.** Gaidos and Nimmo (2000) and Nimmo and Gaidos (2002) suggest that diurnally induced strike-slip motion along fractures creates frictional and viscous heating. If the velocity of motion along a fracture is great enough (∼10 cm per tidal cycle), shear heating can be sufficient to trigger upwelling of warm ice or compression along the weakened zone to form a ridge (Han and Showman, 2008). Shear heating may provide a more uniform mechanism for buoyancy generation that may explain some differences between ridges on Europa and terrestrial linear diapirs. Partial melting might occur along the ridge axis, with downward drainage of melt contributing to formation of the axial depression. On the other hand, neither shear heating nor linear diapirism would explain flexural troughs or tension fractures that flank large ridges.

**Volumetric deformation.** Aydin (2006) notes the similarity in morphology and merging relationships of Europa’s ridges to some compaction and dilation bands in terrestrial rocks, localized zones of volumetric strain that occur in high porosity materials such as sandstone. Moreover, the multiple sets of ridges that comprise Europa’s ridge complexes have a strong resemblance to shear bands in terrestrial rocks. Europa’s ridges and bands are several orders of magnitude larger than the terrestrial analogues formed by volumetric deformation. The potential mechanisms for concentrating strain in large structures along discrete boundaries on Europa need to be understood in order to evaluate the viability of this model.

The origin of ridges, their relationships to isolated troughs and to bands, and their connectedness to the subsurface remain key open issues in our understanding of Europa’s tectonics, but the shear heating model is the most quantitatively developed, and in some ways the most testable model (see also Section 5.2).

5.1.1.4 Bands  Bands are polygonal areas of smoother terrain with sharp boundaries. Bands, like other features on Europa, appear to brighten with age; and the youngest bands are commonly of lower albedo than their surroundings, while older bands show little or no albedo contrast. Opposing sides of bands on Europa can be reconstructed with few gaps, restoring structures that were split and displaced as the bands opened along fractures (Schenk and McKinnon, 1989; Pappalardo and Sullivan, 1996; Sullivan et al., 1998) (Figure 7.9). Reconstruction of bands implies that Europa’s surface layer has behaved in a brittle manner, separating and translating atop a low-viscosity subsurface material, with the region of separation being infilled with relatively dark, mobile material (Schenk and McKinnon, 1989;
Figure 7.9. Prominent band on Europa. Arrows point to some of the preexisting ridges cut by this band. Preexisting features reconstruct almost perfectly if the band is closed, demonstrating that the bands form by crustal spreading. North is to the top, and the illumination is from the right.

Golombek and Banerdt, 1990). Thus, bands offer compelling evidence for warm, mobile material in the shallow Europan subsurface at the time of their formation.

It has generally been inferred that bands have formed in response to tension (Schenk and McKinnon, 1989; Golombek and Banerdt, 1990), with the bands near the anti-Jovian point perhaps forming under nearly isotropic tension (Pieri, 1981), consistent with current-day nonsynchronous rotation stresses west of the anti-Jovian point (see subsection 3.2.2). However, some bands, notably Astypalaea Linea, show a significant strike-slip component, suggesting oblique opening (Tufts et al., 1999). Structural relationships within the anomalous bright band Agenor Linea suggest that it formed by right-lateral strike-slip motion (Prockter et al., 2000a), while an analogous bright band Crick Linea on the sub-Jovian hemisphere may have formed with a component of compression (Greenberg, 2004). Wedge-shaped bands at the anti-Jovian point also show evidence of shear, aligned with the opening direction (Schenk and McKinnon, 1989); such shear is not consistent with isotropic tension or with present-day nonsynchronous rotation. Motion along strike-slip faults can generate “wing cracks” splaying from the region of the fault tip where the surface is being extended. Some bands on Europa appear to have originated as wing cracks associated with larger strike-slip features (Schulson, 2002; Kattenhorn, 2004; Kattenhorn and Marshall, 2006).
Regional-scale Galileo images of pull-apart bands show an overall bilateral symmetry (Sullivan et al., 1998; Prockter et al., 2002). Band margins are generally sharp, and reconstruction suggests that some have opened along preexisting ridges, which may have served as zones of weakness (Prockter et al., 2002), although it is not known whether there is any further relationship between bands and ridges. A narrow central trough is common along bands and is remarkably linear and uniform in width along the length of each band. A hummocky textured zone commonly occurs to either side of this trough. Toward the margins of some bands are regularly spaced subparallel ridges and troughs, which are probably domino-style tilted normal fault blocks of the hummocky unit (Figueroedo and Greeley, 2000; Prockter et al., 2000a).

The units and characteristics of Europan pull-apart bands are analogous to those in terrestrial oceanic-spreading environments (e.g., Macdonald, 1982), suggesting that a spreading analogue may be appropriate (Sullivan et al., 1998; Prockter et al., 2002). The axial trough observed in many bands may be the site of plate separation, the flanking hummocky material may represent cryovolcanic material emplaced symmetrically on either side of the band axis, and the subparallel ridges and troughs are likely analogous to the abyssal hills observed along terrestrial mid-ocean ridges. Examples of contemporaneous three-band junctions have been found, with analogy made to ridge-ridge-ridge triple junctions, suggesting further similarities to terrestrial oceanic spreading processes (Head, 2000; Patterson and Head, 2003). The location and spacing of the ridges and troughs were used by Stempel et al. (2005) to obtain local strain rates of $10^{-15}$–$10^{-12}$ s$^{-1}$ and local stresses of 0.4–2 MPa. These strain rates and stresses are consistent with theoretical models of the formation of narrow band-like rifts (Nimmo, 2004d).

An alternate model for band formation is based on the formation of leads in terrestrial sea ice (Pappalardo and Coon, 1996; Greeley et al., 1998). Greenberg et al. (1998) and Tufts et al. (2000) consider that cyclical tension and compression due to Europa’s diurnal tidal flexing might create bands through a ratcheting process. In this view, cracks open during the tensile phase of the diurnal cycle, allowing water to rise and freeze. These cracks are unable to close completely during the compressional phase due to the addition of the new material; hence, the band widens with time as new material is added. As discussed above in the context of the Greenberg et al. (1998) ridge formation model, this model relies on complete cracking through of Europa’s ice shell to the depth of liquid water below. Experiments with wax analogue models show that cyclic strain on an opening rift zone can form band-like features in a thin brittle layer on top of a ductile substrate (Manga and Sinton, 2004). On a deeper level, one may expect that cyclic ratcheting is the native mechanism of crustal extension on Europa, as opposed to quasi-steady mid-ocean ridge spreading driven by distant subduction on the Earth.
Imaging of several bands indicates that they commonly stand topographically higher than the surrounding ridged plains (Malin and Pieri, 1986; Pappalardo and Sullivan, 1996; Giese et al., 1999; Prockter et al., 2000a; Tufts et al., 2000). This is consistent with emplacement of thermally or compositionally buoyant material, such as ice that is warm and/or clean relative to the cold and/or saltier surrounding lithospheric material (cf. Nimmo et al., 2003a).

5.1.1.5 Folds The means by which Europa accommodates the extensional strain from band formation is not well understood. Prockter and Pappalardo (2000) analyzed high-resolution Galileo images of Europa and presented evidence for regional-scale folds in several locations on Europa, the strongest of which is found in the region of Astypalaea Linea. High-resolution Galileo images reveal subtle shading variations suggestive of folds across Astypalaea Linea roughly perpendicular to its trend, and with a fold wavelength of ≈25 km. Strong corroborative evidence for the fold interpretation comes from small-scale structures along the inferred anticline and syncline axes: discrete sets of small-scale fractures (troughs) occur along the crests of the regional-scale anticlines, while small-scale closely spaced single ridge crests occur within the inferred synclinal lows, trending parallel to the fold axes. They are inferred to be contractional structures (folds and/or thrust blocks) formed within regional-scale synclines.

The Astypalaea folds can be used to constrain the character and thickness of the lithosphere at the time of deformation, and the nature of the stresses that likely formed them. Folds may have formed by means of a compressional instability of a frictionally controlled brittle ice lithosphere overlying a ductile asthenosphere, in which ice strength decreases with depth (Herrick and Stevenson, 1990). If the local thermal gradient is very high (≈100 K km⁻¹) and the brittle lithosphere is correspondingly thin (≈2.5 km), compressional instability can be achieved with approximately 10 MPa of compressional stress, which is a relatively high amount of stress to achieve on Europa (Dombard and McKinnon, 2006b). The high stress derived may be a clue that brittle ice on Europa’s surface is weaker in compression than currently thought.

Other more tentative examples of regional-scale folds have been identified, in the gray band Libya Linea and in the Manannán region (Prockter and Pappalardo, 2000) and in the satellite’s northern leading hemisphere (Figueroedo and Greeley, 2000). Some sets of rounded ridges in the ridged plains may represent small-scale folds (Patel et al., 1999a), but the mechanism of creating such small-wavelength fold structures is unclear. Overall, these folds can accommodate only small amounts of strain. As noted above, possible convergent bands have been identified (Greenberg et al., 2002; Sarid et al., 2002; Greenberg, 2004; Kattenholm and Marshall, 2006;
Patterson et al., 2006), and may help to accommodate the strain from extensional bands.

5.1.2 Nonsynchronous rotation of Europa's ice shell

The smooth, presumably floating, ice shell of Europa is unlikely to have large mass asymmetries, and thus may rotate nonsynchronously. A lower limit of $10^4$ years has been derived for the period of any ongoing nonsynchronous rotation, based on comparison of terminator views of the same features in Voyager 2 and Galileo images obtained 17 years apart (Hoppa et al., 1999c). For their estimated shell thickness $t_c$ of $\sim$15–25 km (from tidal heating calculations), Ojakangas and Stevenson (1989a) predicted a nonsynchronous rotation time of $\sim$10 Myr, consistent with the lower limit of Hoppa et al. (1999c). This timescale goes as $\sqrt{t_c}$; thus, a 10-km thick shell could rotate in 2.5 Myr.

Helfenstein and Parmentier (1985) first predicted the stress pattern that should result from nonsynchronous rotation, based on an eastward shift of Europa’s surface relative to its fixed tidal axes (see subsection 3.2.2). Voyager global-scale lineaments were compared to this pattern by McEwen (1986) and Leith and McKinnon (1996). These workers concluded that the best match of the nonsynchronous stress pattern to Europa’s global-scale lineaments occurred by considering a westward longitudinal shift in the locations of surface features relative to the fixed tidal axes. If the longitude of surface features is shifted westward (or equivalently, the tidal axes shifted eastward) to “back up” nonsynchronous rotation by $\sim$25$^\circ$, then lineament orientations achieve a best fit in being approximately perpendicular to the least compressive (greatest tensile) stress direction, as expected if the lineaments originated as tension fractures. The implication of this best fit is that Europa’s major lineaments may have formed over a range of $\sim$50$^\circ$ of nonsynchronous rotation. Stresses generated by nonsynchronous rotation can be significant. Maximum stresses of $\sim$0.14 MPa can be achieved per degree of rotation (Equation 7.5); thus, accumulated tensile stress can exceed the tensile strength of cold laboratory ice upon $\sim$12$^\circ$ of nonsynchronous rotation (Leith and McKinnon, 1996). However, the proper value of tensile strength to use for ice at the surface of Europa is uncertain due to a paucity of relevant laboratory data (see Section 2.3) and the problems inherent in scaling-up laboratory results to describe the bulk properties of Europa’s lithosphere.

Galileo color images support and strengthen this argument, as imaging at near-infrared wavelengths can discriminate older lineaments that were invisible to Voyager (Clark et al., 1998; Geissler et al., 1998a). Geissler et al. (1998b) categorized lineament age based on color characteristics, and found that lineament orientations have progressively rotated clockwise over time, implying that stress orientation rotated similarly. This rotation sense is just as predicted by eastward migration of
the surface relative to fixed tidal axes due to nonsynchronous rotation. Nonsynchronous rotation is not necessarily the formational stress mechanism, however, as the orientations of the most recent lineaments mapped by Geissler et al. (1998b) are better fit by diurnal stressing. It is plausible that diurnal stressing may create some cracks, while nonsynchronous rotation opens those cracks into wider ridges and bands.

Higher resolution Galileo imaging shows that Europa’s ridged plains are overprinted by ridges and ridge sets of various orientations. Crosscutting relationships inferred from these higher resolution images have been cited as evidence for at least one full rotation of Europa’s ice shell (Geissler et al., 1999; Figueredo and Greeley, 2000; Kattenhorn, 2002). Others have argued from stratigraphic relationships that few structures have formed over each rotation of the ice shell, implying that the surface records several shell rotations (Sarid et al., 2004, 2005), or even hundreds or thousands of shell rotations (Hoppa et al., 2001). Stratigraphic analysis of cycloidal cracks (subsection 5.1.3), which are very sensitive to the ice shell orientation, shows almost two complete rotations are required (Groenleer and Kattenhorn, 2008).

The equatorial region of isotropic tension west of the anti-Jovian point, predicted by nonsynchronous rotation, correlates to the zone of pull-apart bands originally recognized in Voyager imaging (Helfenstein and Parmentier, 1980; Pieri, 1981; Lucchitta and Soderblom, 1982; Schenk and McKinnon, 1989), and recognized from Galileo imaging to extend westward to $\sim 250^\circ$ longitude (Sullivan et al., 1998). A similar extensional region is predicted west of the sub-Jovian point, but is not observed in Galileo hemispheric-scale imaging, perhaps because cracks formed in this region did not open into bands (Hoppa et al., 2000). Zones of contractional tectonics are also predicted, centered $90^\circ$ in longitude away from the extensional zones (Figure 7.4a). Evidence is mounting that shear failure may occur in these zones, as suggested by the “X”-patterned orientations of structures within and just east of these regions (Spaun et al., 2003; Stempel and Pappalardo, 2002).

More complex (but uncertain) stress sources are implied by the findings that the anti-Jovian extensional zone is centered $\sim 15^\circ$ south of the equator (implying that polar wander may have occurred; cf. subsection 3.2.3), and that dark and wedge-shaped band opening directions within have preferred orientations (Schenk and McKinnon, 1989; Sullivan et al., 1998). More evidence for polar wander comes from a survey of strike-slip offsets on Europa, showing that the dividing line between the expected north–south hemispheric dichotomy (see subsection 3.2.1) is tilted $\sim 30^\circ$ relative to the equator (Sarid et al., 2002). Mysterious troughs on Europa that exactly follow antipodal small circles centered near the equator are also offset as if several degrees of polar wander have occurred, and may themselves be the result of an episode of much greater ($\sim 90^\circ$) polar wander (Schenk et al., 2008).
5.1.3 Diurnal tidal variations

Europa is also subject to diurnal tidal variations that impose a daily rotating stress field on the surface, which may change the orientation at which cracks open relative to a larger stress field (e.g., from nonsynchronous rotation), and can lead to ratcheting of strike-slip faults (cf. subsection 3.2.1). Galileo images show many convincing examples of strike-slip offsets on Europa. Hoppa et al. (1999a) note that there is a preferred sense of strike-slip motion in each hemisphere, with a propensity for right-lateral strike-slip faults in the southern hemisphere, and left-lateral in the northern.

Even more definitive evidence for the role of diurnal stress variations is provided by the elegant explanation they provide for the previously mysterious patterns of cycloid ridges (flexus) and other cycloidal structures (Hoppa et al., 1999b) (Figure 7.10). If a fracture propagates across Europa’s surface at an appropriate
speed (about \(3 \text{ km h}^{-1}\)), the stress orientation rotates during a fraction of the
Europan day such that the propagating fracture traces out the curvature of a single
cycloidal arc. Tensile cracks typically propagate faster than this, but the cracks may
propagate incrementally, slowing the overall rate (Lee et al., 2005). The diurnal
stress then drops below the critical value for fracture propagation until the following
orbit, when tensile stress again increases, reinitiating the fracture propagation and
thus generating the next cycloidal arc. The propagation of each succeeding crack
may be aided by tailcracks that form during the strike-slip sliding that follows
the tension in the diurnal cycle (Kattenhorn and Marshall, 2006; Groenleer and
Kattenhorn, 2008).

This model explains several important observable aspects of cycloidal structures.
First, the arcs of an individual cycloidal chain always show a consistent direction
of convexity, while different chains can have opposite convexity directions. In
the diurnal cracking model, convexity direction simply depends of the fracture
propagation direction relative to the sense of stress rotation. Some cycloidal features
transition into linear features, accounted for in the model as fractures propagate into
regions in which the fracture can propagate for only a small fraction of the diurnal
cycle. Similarly, the overall curvature of a cycloidal chain reflects the regional
change in stress orientations from the latitude and longitude regime in which the
fracture initiated, and into which it propagates.

The shape of cycloidal chains on Europa can be closely matched if tensile failure
occurs at a stress of about 25 kPa, if propagation speed drops in proportion to tensile
stress (producing a good match to arc skewness), and propagation halts when
stress drops below 15 kPa. This suggests that Europa’s uppermost lithosphere has a
strength of only \(\sim 25\) kPa. To what depth might these cycloidal fractures penetrate,
and are they expected to transition into normal faults at depth? Leith and McKinnon
(1996) suggest modeling fractures as crevasses, which can extend at least to a depth
\(z = \pi \sigma /2 \rho g\). For an applied tensile stress \(\sigma = 40\) kPa, the corresponding fracture
depth is about 50 m or deeper if the brittle shell is thin (Lee et al., 2005). Thus,
diurnal stressing of a weak Europan lithosphere (tensile strength \(\sigma_0 \sim 25\) kPa) may
produce tensile fractures \(\sim 50\) m to a few hundred meters deep.

5.1.4 Is Europa currently active?

The observed number of impact craters on Europa can be used to estimate the
satellite’s age if accurate estimates of the impactor flux can be made. By modeling
the dynamics of small solar system bodies, Zahnle et al. (1998) and Levison et
al. (2000) conclude that Jupiter family comets are the most common impactors
onto the Galilean satellites in the present epoch. These authors model a current
formation rate of one crater >20 km diameter each 3.2 Myr. The number of observed
large craters on Europa implies a surface age of $\sim 60$ Myr, with a factor of three uncertainty (Zahnle et al., 2003; Schenk et al., 2004b).

An independent method for constraining the age of Europa’s surface comes from estimates of ice sputtering, the ejection of ice particles due to the flux onto Europa of high-energy particles corotating with Jupiter’s magnetic field. Based on Galileo Energetic Particle Detector (EPD) measurements, Ip et al. (2000) and Cooper et al. (2001) each have estimated H$_2$O sputtering rates, with estimates varying from 1.6 to 56 cm Myr$^{-1}$. From high-resolution images, Europa’s stratigraphically oldest units (the ridged plains) display topography on vertical scales of tens of meters. If sputtering rates have been essentially constant over time, then 10 m of topography would be erased in $\sim 2 \times 10^7$ to $6 \times 10^8$ yr, so the oldest regions of Europa must be much less than a billion years old.

If Europa’s average surface age is only $\sim 50$ Myr old, then it seems likely that Europa continues to be tectonically active today. One indication of relatively recent geological activity comes from the photometric properties of some ridge flanks, which suggest immature materials compared to other terrains (Helfenstein et al., 1998), but the aging rate is unknown. As yet there is no certain evidence for current activity on the satellite. Comparison of Voyager and Galileo images obtained at similar resolution and lighting geometries shows no apparent changes, constraining the average surface age to $\geq 30$ Myr (Phillips et al., 2000). Searches for plumes and plume deposits have also been fruitless (Phillips et al., 2000), though the opportunities to search for such activity during the Galileo mission were limited.

### 5.2. Enceladus

Enceladus is one of the most intriguing icy satellites. Despite its small size (252 km radius), Voyager 2 images revealed that portions of its surface are extensively tectonically deformed (Smith et al., 1982; Squyres et al., 1983; Kargel and Pozio, 1996). More recently, Cassini images of the south pole revealed active plumes of water vapor and ice crystals (Porco et al., 2006) emanating from localized thermal anomalies (Spencer et al., 2006) associated with tectonic features referred to as “tiger stripes.” Because Cassini images are still being acquired and interpreted, all conclusions presented here are necessarily preliminary, and tectonics on Enceladus is likely to remain a field of active investigation for some time to come.

The internal structure of Enceladus probably consists of an ice shell 100 km or so thick overlying a silicate core (e.g., Schubert et al., 2007). Because of the tectonic deformation and current activity observed at the surface, there is likely a subsurface ocean that decouples the ice shell from the silicate interior and increases the tidal deformation. Based on shape data, Enceladus probably is not in global
hydrostatic equilibrium (Thomas et al., 2007), which makes determination of its moment of inertia and interior structure very difficult.

Tidal stresses are likely an important source of deformation on Enceladus. If the shell of Enceladus responded in a fluid fashion, then the diurnal stresses and strain rates would be factors of 6 and 15 larger than Europa, respectively (Nimmo et al., 2007b; Table 7.1). In practice, the ice shell of Enceladus is likely to be sufficiently thick that the deformation is reduced (Ross and Schubert, 1989); nonetheless, the stresses generated can be significant. A rough estimate of their magnitude is $250 h_2$ kPa, where the Love number $h_2$ gives the response of the surface to tides and is 2.5 for a homogeneous fluid body (Nimmo et al., 2007b).

Voyager 2 encountered Enceladus during northern summer, and thus observations were concentrated on the northern hemisphere. Cassini observations to date have been complementary to the Voyager observations, as they were obtained during southern summer and mainly focused on the south polar region. Most of the satellite (except for a swath centered on roughly 80°W) has now been imaged well enough to carry out global tectonic mapping. Broadly speaking, the satellite may be divided into three terrains: older cratered terrain, tectonically disrupted terrain, and the south polar terrain. Cratered terrains (Figure 7.11) occupy a broad band encircling the satellite along the 0° and 180° longitude lines (over the north pole, through the sub- and antisaturnian points). The craters in the older terrain are in some cases superimposed on sets of ancient, even older linear structures, and are often cut by younger fractures. In high-resolution images, many of the recent fractures in the cratered terrain are observed to be chains of pits (Michaud et al., 2008), which are most likely formed by regolith drainage over dilational normal faults (e.g., Wyrick et al., 2004) with a small amount of extensional strain.

Centered on the leading and trailing hemispheres (90° and 270°) are younger, tectonically disrupted terrains roughly 90° wide at the equator. This terrain is characterized by densely spaced sets of subparallel ridges and troughs (Figure 7.11), reminiscent of grooved terrain on Ganymede (subsection 6.1.2), and is probably the result of normal faulting. This terrain also harbors some prominent rounded, branching ridges that rise well above the surrounding terrain, the origin of which is not well understood.

Although the north polar region is apparently undeformed cratered terrain, the south polar region (south of 55°S) is complexly tectonized and essentially uncratered (Figure 7.12; Porco et al., 2006). The margins of this region are scalloped, and cusps that extend northward into the surrounding regions contain arcuate scarps that are likely fold-and-thrust belts (Helfenstein et al., 2006a). North–south trending extensional-tectonic structures stretch towards the equator from the ends of the cusps. Within the south polar terrain, the most prominent features are a set of subparallel, ridge-flanked troughs up to 0.5 km deep, 2 km wide and $\sim$130 km
long, informally termed “tiger stripes” (Porco et al., 2006). These troughs trend at about 45° from the tidal axis of Enceladus’ figure, and have a spacing of about 35 km.

A broad range of crater densities confirms that Enceladus has had a long and at least intermittently active tectonic history (Smith et al., 1982; Kargel and Pozio, 1996; Porco et al., 2006). The most heavily cratered regions have impact crater densities, suitably scaled, that are comparable to those of the lunar highlands (Porco et al., 2006) and suggest that some terrains have survived mostly intact for billions of years. On the other hand, some of the tectonically deformed regions possess very few craters, indicating that these terrains were deformed within the last few

Figure 7.11. Cratered terrain (right) cut by a swath of tectonized terrain (left) near the equator of Saturn’s satellite Enceladus. North is to the top, and the illumination is from the left.
Planetary Tectonics

Figure 7.12. The south polar terrain on Enceladus forms a complexly tectonized area bounded by a band of cuspatate ridges. Bands of extensional faults tend to radiate toward the equator from the outermost projections of the cuspatate boundary. The “tiger stripes” are the dark ridge pairs in the center of the image. Polar stereographic map projection, centered on the south pole. The blurry area in the upper left quadrant has not been imaged at high resolution at the time of writing.

tens of millions of years, and the south polar region is devoid of craters \( > 1 \text{ km} \) diameter, indicating a surface age of \( \sim 1 \text{ Myr} \) (Porco et al., 2006). In some cases, the observed craters are much shallower than would be expected, indicating that relaxation has occurred, presumably as a result of relatively high subsurface heat fluxes (Passey, 1983; Schenk and Moore, 1995; Porco et al., 2006).

Because of its intense geological activity and low surface gravity, Enceladus is topographically relatively rough. Limb profiles can in some cases be correlated with geological features such as Samarkand Sulci (one of the equatorial bands
of deformed terrain), and reveal local relief of up to 1 km vertical over \(\sim 100 \text{ km}\) distance (Kargel and Pozio, 1996 cf. Bland et al., 2007). Certain trough flanks are suggestive of flank uplift, implying flexural parameters of roughly 30 km and indicating elastic thicknesses of about 4 km (Kargel and Pozio, 1996). The south polar region is depressed relative to the surrounding terrain by \(\sim 0.5 \text{ km}\) (Thomas et al., 2007), possibly as a result of melting of ice by a subsurface heat source, creating a trapped sea beneath the south pole (Collins and Goodman, 2007).

Our understanding of the geological evolution of Enceladus, and the mechanisms responsible, is still evolving. The wide range of surface ages suggests patchy resurfacing, perhaps in a somewhat analogous manner to Ganymede (Bland et al., 2007). Based on Voyager 2 images, cryovolcanism was favored as the principal resurfacing mechanism (Squyres et al., 1983), but Cassini images suggest that this resurfacing is primarily tectonic (Porco et al., 2006; Barr, 2008). In addition, fallout from the plumes may cause mantling of surface features, and localized cryovolcanism has been suggested.

Most Cassini-based work to date has focused on the evolution of the south polar region. The orientation of the tiger stripes is consistent with their original formation as tension cracks, because they are oriented perpendicular to the maximum present-day diurnal stresses (Nimmo et al., 2007b). The tiger stripes may currently be undergoing strike-slip motion, possibly explaining the existence and timing of the plume eruptions (Nimmo et al., 2007b; Hurford et al., 2007; Smith-Konter and Pappalardo, 2008). No evidence has yet been reported for geological strike-slip offsets, but the deformation is expected to be cyclic. The existence of the unusually high heat flow must ultimately be due to tidal heating in Enceladus’ south pole region, and may be tied to either the presence of a subsurface warm diapir (Nimmo and Pappalardo, 2006) or to melting and subsidence associated with heating in the silicate core or in the ice shell (Collins and Goodman, 2007; Tobie et al., 2008). Both of these mechanisms are likely to produce gravity anomalies leading to poleward reorientation of the region, thus explaining its current polar location (Nimmo and Pappalardo, 2006), although tidal heating should theoretically be maximized at the poles in any case. They also lead to predictable local tectonic stresses: extensional for a rising diapir, and compressional for a subsiding surface (see Section 6.2 for a similar situation on Miranda).

Poleward reorientation in turn generates its own set of global stresses (e.g., Melosh, 1980a; Leith and McKinnon, 1996; subsection 3.2.3). For synchronous satellites, reorientation is complicated by the triaxial shape of the body, with the result that reorientation tends to happen most readily around the tidal axis (e.g., Matsuyama and Nimmo, 2007). The resulting stresses can be calculated (Leith and McKinnon, 1996) and compared with existing tectonic features. Whether such reorientation stresses can explain the tectonic features observed is currently a topic
of investigation. The identification of potential “tiger stripe analogues” in near-equatorial regions on Enceladus (Helfenstein et al., 2006b) is a potential argument for reorientation, but this is also a topic of continuing research.

Another potentially important source of stress arises from an ice shell thickening above a subsurface ocean (Nimmo, 2004b). Because ice expands as it freezes, the surface moves radially outwards, leading to isotropic extension. Thus, the predominance of extensional features on Enceladus may be a function of its ice shell having thickened with time. Ice shell freezing can also lead to pressurization of the underlying ocean, and potentially cryovolcanism (Manga and Wang, 2007).

In summary, Enceladus is a geologically active, heavily deformed satellite that has a visible geological history stretching from billions of years ago to the present day. Our understanding is currently at a crude level. With further Cassini flybys planned and intensive data analysis only now beginning, the unraveling of this fascinating body’s tectonic history has just begun.

5.3 Triton

Triton is the only large satellite of Neptune, and has a mass ∼40% greater than Pluto’s. As Voyager 2 flew by in 1989, slightly more than half the surface was imaged and ∼20% of that was obtained at resolutions of 1 km/pixel or better, sufficient to discriminate geological features at regional scales. These images revealed a young surface with relatively few impact craters and a wealth of geological landforms. Some are interpreted to be cryovolcanic in origin, including flow lobes and pyroclastic sheets, and some are likely to have a tectonic origin, such as ridges and troughs. The geological age of the portion of Triton’s surface imaged by Voyager is of the order 100 Myr (Stern and McKinnon, 2000; Schenk and Zahnle, 2007), implying that it follows Io, Europa, and portions of Enceladus in its level of geological activity. The dominant “bedrock” material is thought to be water ice or ammonia hydrate ice (Croft et al., 1995; Cruikshank et al., 2000), although a number of other exotic ices have been observed (e.g., Na, CO$_2$; Quirico et al., 1999). Tidal heating may have sustained warm interior temperatures for upwards of a billion years, and if ammonia is present in the icy mantle, a subsurface liquid ocean may still persist today (Hussmann et al., 2006).

Triton is in a retrograde and highly inclined (157°) orbit around Neptune, suggesting that it did not originate in its current position but is instead a captured object. Several models for Triton’s capture have been proposed, including aerodynamic drag in a protosatellite disk (McKinnon and Leith, 1995), capture by collision with an existing satellite (Goldreich et al., 1989; Benner and McKinnon, 1995), or exchange capture between a binary system and Neptune, in which one member of the binary was expelled and its place taken by the planet (Agnor and...
Figure 7.13. Neptune’s satellite Triton exhibits several bands of sinuous ridges and double-ridge features reminiscent of those on Europa. This image shows the intersection of Slidr Sulci (running northwest to southeast) with a few smaller orthogonal ridge sets. The terrain surrounding the ridge sets is termed “cantaloupe terrain,” composed of shallow quasi-circular depressions bounded by single ridges. North is to the top, and the illumination is from the bottom.

Hamilton, 2006). Of these, aerodynamic drag is expected to have occurred early in Neptune’s history within a specific time period. Conditions for collisional and exchange capture are less time-sensitive, as Neptune migrated outwards due to encounters with material from the protoplanetary disk. Following capture, the orbit is thought to have circularized on a timescale of several hundred Myr to \( \sim 1 \) Gyr (Ross and Schubert, 1990; McKinnon et al., 1995).

5.3.1. Ridges

Triton has a significant number of sinuous ridges (Figure 7.13), spanning a range of ages and degradation states. Typical ridges measure \( \sim 15–20 \) km in width, and some are continuous for \( \sim 800 \) km. A deep continuous axial depression \( \sim 5–10 \) km wide flanked by higher ridges results in a double-ridge morphology, and some ridges are bounded by shallow troughs \( \sim 20 \) km wide. Triton is the only other place in the solar system where Europa-like double ridges have been identified (subsection 5.1.1.3), and indeed these appear to be the dominant tectonic structure on Triton. The ridges on Triton are remarkably similar to Europan ridges, including single isolated troughs and double ridges, as well as much rarer triple and multi-crested ridges (Prockter et al., 2005). Triton’s ridges are typically many times wider than typical Europan ridges, but are morphologically subdued – the very limited available topography (Croft et al., 1995) suggests that they are only a couple of hundred meters high, similar to Europan ridges.
Voyager-based models for ridge formation on Triton suggested cryovolcanic extrusion into graben (Smith et al., 1989; Croft et al., 1995). Investigations into ridge formation on Europa suggest that the graben model does not fit the observed morphologies, and given the strong resemblance of ridges on the two moons, it is reasonable that similar formation mechanisms are responsible on both. The shear heating model of ridge formation (Gaidos and Nimmo, 2000; Nimmo and Gaidos, 2002) was proposed for Triton’s ridges by Prockter et al. (2005). They found that the magnitudes of stresses and heat fluxes required to generate ridges of the correct scale are comparable to predicted values generated during Triton’s orbital evolution from a highly eccentric state. The much greater widths of Triton’s ridges compared to ridges on Europa are likely due to the lower surface temperature, and thus a greater brittle–ductile transition depth.

The large-scale pattern of ridges and troughs on Triton is still not well understood. Equivocal results have come from attempts to compare the lineament patterns to plausible global stress models including despinning, orbital precession, and non-synchronous rotation (Collins and Schenk, 1994; Croft et al., 1995; Prockter et al., 2005).

If Triton’s ridges do form by shear heating, the timescale becomes puzzling. Heat generated during capture could have easily melted Triton’s interior (McKinnon et al., 1995), probably enabling the surface to rapidly and completely overturn. Could the current surface be a relic of Triton’s waning tidally driven, geological activity? The ridges may not themselves be young, but the lack of impact craters and otherwise young surface suggests they formed relatively recently, which implies that capture was also a relatively recent event. Because the timescale for Triton’s orbital evolution depends on the poorly known (and time-variable) internal mechanical layering of Triton and Neptune, the time of Triton’s capture is not well constrained.

5.3.2. Tectonic interactions with cryovolcanic deposits

The surface of Triton exhibits many landforms that resemble terrestrial volcanic features (Croft et al., 1995). These include cones 7–15 km in diameter, occurring individually or in clusters; chains of pits along ridges are similar to terrestrial tectonically controlled cinder cones and explosion pits; and circular to elongate depressions or pit paterae, occurring singly or in chains and ~10–20 km in diameter, located in patches of smooth material 100–200 km in extent (Figure 7.14). These tend to follow regional tectonic trends and may be analogous to chains of explosion and collapse pits in terrestrial volcano–tectonic zones. Some of the most enigmatic features are the larger ring paterae, 50–100 km in scale, with an outer rim defined by a ring of coalescing pits. The pit and ring paterae along with their associated
smooth deposits are interpreted to have formed as explosive cryovolcanic craters and deposits.

Triton’s enigmatic “cantaloupe terrain” contains quasi-circular shallow depressions termed cavi, typically 25–35 km in diameter with slightly raised rims, giving the appearance of a cantaloupe rind (Figure 7.13; Croft et al., 1995). Cavi have been suggested to represent cryovolcanic explosion craters, such as terrestrial maars, on the basis of their similar morphologies. However, the organized cellular pattern of the cantaloupe terrain has been proposed to closely resemble the expression of terrestrial salt diapirs, and the terrain may have formed due to diapirism resulting from gravity-driven overturn within an ice crust about 20 km thick (Schenk and Jackson, 1993). One possibility is that cryovolcanism on Triton layered more dense ices on top of less dense ice layers, leading to compositionally driven overturn of the ice layers. Cavi may be analogous to similar structures on Europa, which have also been proposed to have a diapiric origin.

5.3.3. Current activity

Voyager images of Triton’s bright south polar cap revealed dark streaks, possibly methane converted to organic material by energetic particles and ultraviolet photons. This observation was remarkable given that Triton is thought to undergo a cycle of volatile deposition and sublimation from pole to pole on the cycle of 1 Triton year, or ~165 Earth years (Buratti et al., 1994). This yearly cycle is expected to result in a meter or more of nitrogen, methane, and carbon dioxide frosts being sublimated from one pole and deposited on the other, thus the presence of the dark streaks implied they were very young. Searches of stereoscopic images of the south polar region clearly show plumes, up to 8 km in height with radii of up to 1 km,
probably composed of dust and gas (Kirk et al., 1995). The columns feed clouds of
dark material that drifted with Triton’s tenuous winds for more than one hundred
kilometers.

Two models have been proposed to drive the plumes. One suggests explosive
venting of nitrogen gas pressurized by solar heating (Kirk et al., 1995). Triton’s
38 K surface might be blanketed with transparent solid nitrogen. In this model,
dark material lying immediately below the transparent layer is warmed by sunlight,
undergoing a significant increase in temperature with respect to the surface and a
corresponding increase in vapor pressure of the surrounding nitrogen. The highly
pressurized nitrogen is trapped in pore spaces, then released to the surface through
a vent, entraining dark material and lofting it into the atmosphere. The model
suggests that a temperature increase of only 2° would be sufficient to propel the
plumes to the observed altitudes. An alternative driving mechanism suggests that
the heat source for the geysers comes from within the satellite, perhaps due to
thermal convection in the underlying ice (Duxbury and Brown, 1997). This model
seems consistent with Triton’s young surface age and the extreme tidal heating
predicted during capture of the satellite by Neptune.

6 Formerly active icy satellites
6.1 Ganymede

Ganymede is the largest satellite in the solar system, larger than the planet Mercury.
It has the lowest normalized axial moment of inertia of any solid body known in the
solar system, indicating that its mass is highly concentrated in the interior. Probably
its interior consists of an iron core several hundred kilometers in radius, followed
by a rocky mantle, with about 800 km of water ice on top (Anderson et al., 1996).
A molten iron core likely explains the internally generated magnetic dipole of
Ganymede (Schubert et al., 1996), the only satellite with its own magnetosphere.
Variations in Ganymede’s magnetic field also point to the existence of a subsur-
face ocean approximately 150 km below the surface, sandwiched between low
pressure ice I at the surface and higher pressure phases of ice below (Kivelson
et al., 2002). The majority of its surface is dominated by tectonic features, but
even the youngest of these features is overlain by younger impact craters, and is
nominally 2 billion years old (Zahnle et al., 2003), though there is considerable
uncertainty in that age estimate. The surface of Ganymede is commonly divided
into two broad categories, termed dark terrain and bright terrain. We will discuss
the tectonics on both of these terrains, and then summarize the implications for
tectonic driving mechanisms on Ganymede (see Pappalardo et al. (2004) for an
up-to-date summary of all geology on Ganymede).
Figure 7.15. Ancient structures known as “furrows” arc from northwest to southeast in this area of Galileo Regio on Ganymede. North is to the top, and the illumination is from the left.

6.1.1. Dark terrain

Dark terrain covers one third of the surface of Ganymede and appears to be an ancient, perhaps primordial, surface. The surface is saturated with impact craters, and a surface layer of loose dark dust mantles much of the topography (Prockter et al., 1998). Viewed from a distance, many areas of dark terrain are dominated by sets of concentric ring arcs, termed furrows (Figure 7.15). Furrow sets can be thousands of kilometers across and are interpreted to be concentric fractures around ancient impact basins. As such, they are not endogenic tectonic features, but their characteristics do tell us something about the nature of the lithosphere in which they formed. Furrows must have formed in a lithosphere that was relatively thin (McKinnon and Melosh, 1980) compared to the present day, since more recent large impacts such as the basin Gilgamesh did not form these closely spaced features as they collapsed. Furrows are the oldest recognized feature on Ganymede, being cut by all other craters (Passey and Shoemaker, 1982), and thus giving us insight
into an early period of higher heat flow and thinner lithosphere. The furrows themselves consist of two parallel ridges with a trough in between. They are interpreted to be graben-like features that have undergone topographic relaxation (McKinnon and Melosh, 1980). Nimmo and Pappalardo (2004) fit flexural models to the rift flank uplift on the furrow features and estimate that an ancient heat flux of 60–80 mW m\(^{-2}\) is necessary to fit the topography, corresponding to an elastic lithosphere 2–3 km thick. It is unknown, though, whether this represents the heat flux during the formation of the furrows or during the formation of bright terrain (see subsection 6.1.2), and how much the topography has viscously relaxed since the furrows formed.

From a distance, dark terrain also exhibits higher albedo streaks or irregular patches. Close-up Galileo images have resolved these to be concentrated areas of intense faulting. Motion along faults exposes bright ice below the dark surface, brightening the terrain. This may be one mode by which dark terrain changes into bright terrain (Prockter et al., 2000b). In some areas of dark terrain, it appears that recent deformation has concentrated in former furrows, perhaps utilizing them as preexisting weaknesses in the lithosphere (Murchie et al., 1986).

### 6.1.2 Bright terrain

The distinctive appearance of Ganymede results from the bright terrain that slices the dark terrain into discrete polygons. Within the bright terrain is a mosaic of crosscutting swaths and polygons of subparallel ridges and troughs (termed grooved terrain) or smooth bright plains (Figure 7.16). Grooved terrain exhibits several different morphologies of ridges and troughs, which appear to be related to the amount of strain accommodated in each polygon of the bright terrain. So far, no terrain with evidence for contractional strain has been identified. There is, however, abundant evidence for extensional strain all over the surface of Ganymede.

At the low-strain end of the strain spectrum are bright plains cut by parallel dark troughs and graben-like structures (Figure 7.17). Measurement of craters cut by grooves of this morphology show less than 5% extensional strain (Pappalardo and Collins, 2005). The origin of the smooth bright plains is unclear, though it has been suggested that they represent areas flooded by cryovolcanic flows (e.g., Shoemaker et al., 1982), which is supported by the even, low topography of some of these areas (Schenk et al., 2001b). However, no bright plains have been found without some form of ridges or troughs on their surfaces, so tectonism appears to be an essential part of bright terrain formation.

At the other end of the spectrum are areas of grooved terrain that appear to exhibit tilt-block normal faulting (Pappalardo et al., 1998b). High-resolution observations of these areas show a landscape of parallel ridges and troughs with a triangular
Figure 7.16. Broad-scale view of Ganymede grooved terrain, in Nun Sulci on the sub-Jovian hemisphere. North is to the top, and the illumination is from the left.

Figure 7.17. High-resolution view of graben sets in grooved terrain, Uruk Sulcus, Ganymede. North is to the top.
sawtooth cross section (Figure 7.18). In the tilt-block normal faulting model, triangular ridges are formed as the surface is cut apart by parallel normal faults, all dipping in the same direction. As the terrain is pulled apart, motion along the fault exposes the fault scarp on one side of the ridge, while the original surface tilts back to form the other side of the ridge. These tilt-block ridges on Ganymede are typically ~1 km wide and 100–200 m high.

The geometry of this tilt-block faulting model was used by Collins et al. (1998a) to estimate that one region of grooved terrain had been pulled apart by about 50%, based on the throw on the fault scarps. In three other areas, Pappalardo and Collins (2005) found impact craters that had been cut by tilt-block faulting zones about 10–20 km in width. Using the craters as strain markers, they estimated that these sets of faults had accommodated from 50% up to 180% extensional strain. Some of these sets of faults had also accommodated a few kilometers of strike-slip motion in addition to the extension normal to the faults. The strain is high enough that no features from the preexisting surface can be recognized within the fault zone, a process termed tectonic resurfacing, which can wipe out craters and reset the surface age through tectonism alone.

In addition to extensional deformation, strike-slip motion is also observed as a component of Ganymede tectonics. Strike-slip motion was suspected from Voyager
data based on the sigmoidal shapes of many small regions of grooved terrain, and
offsets of background terrain features on either side of grooved terrain swaths
(Lucchitta, 1980; Murchie and Head, 1988). Higher resolution observations show
several fault zones in which normal faults are organized into en echelon segments,
indicating transtension (Pappalardo et al., 1998b; Collins et al., 1998b; DeRemer
and Pappalardo, 2003). Two of the five fault zones measured by Pappalardo and
Collins (2005) using craters as strain markers show significant levels of strike-slip
motion along the faults.

In the tilt-block regions, the small-scale ridges and troughs described above
are superimposed on broader undulating topography, with a ridge spacing (or
wavelength) of \( \sim 5 \) to 10 km (Patel et al., 1999b) and an overall relief of about
500 m (Giese et al., 1998; Squyres, 1981). This longer length scale of periodic
deformation may be evidence of extensional instability of the brittle lithosphere
over a ductile substrate (Fink and Fletcher, 1981; Herrick and Stevenson, 1990;
Collins et al., 1998a; Dombard and McKinnon, 2001). In such an instability, the
lithosphere develops periodic pinches where extensional strain is preferentially
concentrated; indeed, in these broad-scale valleys, the tilt blocks become smaller
and may exhibit secondary faulting and greater strain. The existence of these
lithospheric necks helps to constrain the properties of the lithosphere, since they
are sensitive to the thickness of the lithosphere and the ductile properties of the
substance. Using this relationship, Dombard and McKinnon (2001) estimated that
the grooved terrain formed at a strain rate of \( 10^{-16} \) to \( 10^{-14} \) s\(^{-1}\) and a heat flux
of 60–120 mW m\(^{-2}\). However, more recent numerical simulations of extensional
instabilities, using a different lithospheric strength envelope, have found that it may
be more difficult to produce the observed ridges than the analytical models predict
(Bland and Showman, 2007). More robust estimates should result from improved
numerical models (e.g. Bland et al., 2008a).

Using rift flank uplift at the edges of grooved terrain, Nimmo et al. (2002)
estimated a heat flux of 100 mW m\(^{-2}\), in good agreement with the extensional
instability model. The heat fluxes from both these models indicate that grooved
terrain formed with an elastic lithosphere about 1–2 km thick. In order for the
observed faults to undergo shear failure through this lithosphere, stresses on the
order of 1 MPa are required; the possible sources of this stress are the subject of
the next section.

On Europa, smooth bands on the surface appear to be the result of complete
spreading of the lithosphere (see subsection 5.1.1.3). Ganymede exhibits a few
morphologically similar features, notably a 25-km wide swath of bright smooth
material called Arbela Sulcus, cutting across the dark terrain of Nicholson Regio.
Like the bands on Europa, this band can also be reconstructed by rotating one edge
around a pole of rotation, matching preexisting features on the two sides (Head
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et al., 2002). Though this feature is intriguing, it is still unclear how widespread lithospheric spreading is on Ganymede.

6.1.3 Implications for Ganymede evolution

Grooved terrain is the record of an active episode somewhere in the middle of Ganymede’s history, but what could have triggered it? Any hypothesis for the driving mechanism behind groove formation must explain: (a) why it happened in the middle of Ganymede’s history and not near the beginning, (b) how a sufficiently high amount of stress was generated to initiate fault failure, (c) the global, interconnected nature of grooved terrain faults, (d) the high levels of extensional strain observed, with no evidence found yet for contraction, and (e) the high heat flow inferred during the period of groove formation. It would be beneficial if the hypothesis also explained or related to other aspects of Ganymede, such as its molten iron core (e.g., Bland et al., 2008b), and the distribution of craters on its surface.

Most of these points can be addressed by positing a heat pulse in Ganymede’s interior. This could be the result of converting potential energy into heat as the satellite differentiates, or it could be due to enhanced tidal heating at some point in Ganymede’s past. The three innermost Galilean satellites are currently locked in a 4:2:1 orbital resonance, called the Laplace resonance. As the satellites evolved towards this resonance, they may have passed through a Laplace-like resonance, which would have pumped up Ganymede’s orbital eccentricity and caused enhanced tidal heating (Showman et al., 1997). This tidal heating episode is sufficient to explain the inferred heat fluxes, and could also have warmed and melted some of the ice within Ganymede, generating a small amount of volume expansion (Showman et al., 1997), which, in turn, caused surface extension. The tidal resonance hypothesis provides a natural explanation for how to delay the formation of grooved terrain.

If the heat pulse is due to differentiation, there is some question as to how it would remain undifferentiated for the initial part of its history (Friedson and Stevenson, 1983), although the mostly undifferentiated state of Callisto today (Anderson et al., 2001) suggests this is a possibility. Differentiation, or completion of differentiation, could itself have been triggered by the tidal heating episode discussed above. One attractive aspect of differentiation is that since Ganymede is so large, there would be a large volume of high-pressure ice phases in the interior displaced to the outside, which would cause significant volume change (Squyres, 1980; Mueller and McKinnon, 1988). Volume change would cause isotropic tensile stress over the entire surface (see subsection 3.2.6), which may explain the large amount of extensional strain observed. Initial estimates of the amount of global expansion represented by all grooved terrain on Ganymede is significantly higher than would
be expected from heating and melting alone, and is closer to the amount expected
from interior differentiation (Collins, 2006).

Nonsynchronous rotation (subsection 3.2.2) is another possible source of stress
for Ganymede grooved terrain, though the dominance of extension of Ganymede’s
surface argues that it may not be the primary source. The primary evidence for non-
synchronous rotation having occurred on Ganymede comes from its crater popula-
tion. Synchronously rotating satellites accumulate more craters on the hemisphere
that leads their orbital motion, but Ganymede’s younger bright terrain shows a
much weaker than expected asymmetry in crater density (Zahnle et al., 2001).
The simplest explanation for this discrepancy is that Ganymede rotated nonsyn-
chronously for a significant period of time, resulting in a uniform crater population.
Later impacts contributed an anisotropic population, resulting in the diluted pat-
tern we see today. Also, features called catenae on Ganymede are believed to be
formed by comets split into fragments by a close pass by Jupiter (much like comet
Shoemaker-Levy 9). All catenae should form on the Jupiter-facing hemisphere as
the recently split comets leave the Jupiter system (Schenk et al., 1996), but on
Ganymede a few of the catenae are found on the opposite hemisphere, indicating
that it may have faced toward Jupiter for a finite time (Zahnle et al., 2001). True
polar wander has also been proposed on Ganymede (Murchie and Head, 1986;
Mohit et al., 2004), which could also explain the origin of the stresses and the
crater asymmetry, but this finding has not been corroborated.

Ultimately, to distinguish among the hypotheses for grooved terrain formation,
we need to compare the theoretical predictions to the details of the record of
deformation in grooved terrain itself, which in turn will require careful mapping
of the surface. The sparse nature of high-resolution data and the gaps that still
remain in surface coverage mean that a definitive answer may have to wait for a
new mission to the Jupiter system.

6.2 Miranda

Miranda is the smallest of the five major satellites of Uranus, with an average radius
of only 236 km. The surface of Miranda (Figure 7.19) consists of cratered terrain
crosscut by three “coronae,” which are ovoidal to trapezoidal regions of low crater
density, containing sets of ridges and troughs (Smith et al., 1986). The cratered
terrain (Plescia, 1988; Stooke, 1991; Croft and Soderblom, 1991) is characterized
by rolling topography punctuated by large muted craters and smaller sharp ones.
Sharp and muted scarps and inward-facing scarp pairs, interpreted as normal fault
scarps and graben, occur within the cratered terrain. The underpopulation of small
craters and the muting of larger craters and scarps might be due to a large-scale
mantling event, and dark material exposed in the walls of some fresh scarps and
craters also argues for mantling (Croft and Soderblom, 1991). The presence of both muted (pre-mantling) and sharp (post-mantling) fault scarps indicates ongoing or multiple episodes of extensional tectonism during Miranda’s period of endogenic activity. Crater counts suggest that Miranda’s coronae may have been geologically active less than a billion years ago (Zahnle et al., 2003).

The coronae are each comprised of an inner core and an outer belt, as defined by albedo and topographic variations (Smith et al., 1986; Greenberg et al., 1991; Pappalardo, 1994). The inner regions contain smooth material and/or intersecting ridges and troughs; the outer belts are predominantly comprised of distinct bands of subparallel ridges and troughs. Peculiarly, the coronae are “squared off,” with relatively straight boundaries and rounded corners, reminiscent of a race track. The crater population and sharpness of topography within Miranda’s coronae argues that corona formation postdated the event(s) that muted Miranda’s large-scale topography (Plescia, 1988; Croft and Soderblom, 1991; Greenberg et al., 1991).

Upon the encounter of the Voyager spacecraft with Miranda the coronae were interpreted as manifestations of breakup by catastrophic impact and reaccretion of a partially differentiated proto-Miranda (Smith et al., 1986). Janes and Melosh (1988) developed this idea into the “sinker” model, in which silicate-rich chunks of a shattered proto-Miranda sank toward the center of the reaccreting satellite. This would induce a downwelling wake, compressing the lithosphere above. Modeling by Janes and Melosh shows that resulting stresses would create a region of folds.

Figure 7.19. Broad-scale view of Miranda’s surface, showing cratered terrain cut by the angular sets of faults known as coronae. The view is centered near the south pole.
and/or thrusts with no preferred orientation, surrounded by an annulus of concentric folds. This tectonic pattern of a broad load on a small planetary body is distinct from that formed by a small load on a large planetary body, in that satellite curvature and membrane stress is an important factor in the tectonics of Miranda’s coronae (Janes and Melosh, 1990).

Subsequently, a diapiric upwelling or “riser” model was proposed (Croft and Soderblom, 1991; Greenberg et al., 1991; cf. McKinnon, 1988), in which coronae and their constituent ridges and troughs are surface manifestations of large-scale upwelling, perhaps associated with partial differentiation of Miranda. In this scenario, diapirs might pierce and replace the original surface to create coronae or might modify the original surface through extensional-tectonic deformation and extrusion (Greenberg et al., 1991). In predicting the structures formed above a region of upwelling, this model simply requires a sign change from that of Janes and Melosh (1988), predicting a central region of disorganized extensional structures surrounded by a zone of concentric extensional faults (McKinnon, 1988; Janes and Melosh, 1990).

The origin of ridges and troughs within coronae is the principal constraint on models for the formation of coronae. If coronae were formed by downwelling currents, the ridge and trough terrain of their outer belts should be compressional in origin, expressed as folds or reverse faults (Janes and Melosh, 1988). If coronae were formed by upwelling, Miranda’s ridge and trough terrain is predicted to be of extensional–tectonic origin (McKinnon, 1988; Janes and Melosh, 1990; Greenberg et al., 1991), expressed as horst-and-graben structures or tilt blocks, potentially in combination with constructional fissure volcanism and/or intrusion.

A volcano-tectonic model for the evolution of coronae and their constituent ridges and troughs (Croft and Soderblom, 1991; Greenberg et al., 1991) suggests that melt delivered to Miranda’s subsurface erupted through preexisting fractures to create coronae. Consistent with the riser model, this suggests that many of the ridges and troughs within coronae formed by extrusion of viscous material along fissures. Schenk (1991) similarly concludes that many ridges within Elsinore and Inverness Coronae originated by linear extrusion of viscous volcanic material, while Jankowski and Squyres (1988) suggest that some ridges formed by solid-state emplacement of diapiric material.

The morphologies of scarps in Arden and Inverness Coronae, including limb profiles that show asymmetric steps (Figure 7.20), indicates that they were likely formed by normal faulting (Plescia, 1988; Thomas, 1988; Greenberg et al., 1991; Pappalardo, 1994; Pappalardo et al., 1997). Reconstruction of apparent tilt-block style normal faults in the outer belt of Arden Corona suggests that tens of percent extension has occurred, along faults with initial dips of \(\sim 50^\circ\) (Pappalardo et al., 1997).
Figure 7.20. View of the Uranian satellite Miranda, showing normal fault scarps. The faults are formed in parallel sets of tilt blocks, and the sawtooth profile of one set of tilt blocks can be seen on the upper limb.

The weight of morphological evidence suggests that Miranda’s coronae formed by extension and associated cryovolcanism, consistent with a riser model of corona formation. An upwelling origin of Miranda’s coronae eliminates the need to invoke catastrophic breakup and reaccretion of the satellite as an explanation for its surface geology. Instead, coronae may represent the surface expression of broad diapirs rising within a relatively small satellite. The reason for the relatively straight sides and rounded corners of the coronae remains a mystery, but may be the result of structural control by more ancient structures. Miranda’s relatively high current-day inclination is convincingly explained by passage through a tidal resonance with Umbriel, and passage through temporary resonances with Ariel and/or Umbriel would have tidally heated the moon to some degree (e.g., Dermott et al., 1988; Tittemore and Wisdom, 1990; Moons and Henrard, 1994; Peale, 1999). Perhaps it is this episode of tidal heating that is responsible for the renewal of tectonic activity we see on this tiny satellite.
6.3 Ariel

The surface of Ariel can be divided into three geologic units: cratered terrain, presumably the oldest material; smooth plains, displaying a relatively low crater density; and ridged terrain, characterized by bands of subparallel ridges and troughs (Plescia, 1987). Ridged terrain consists of 25- to 70-km wide swaths of east–west or northeast–southwest trending groups of parallel ridges and troughs. The ridges and troughs typically have spacing distances of 10–35 km and can be more than 100 km in length. Smaller scale ridges and troughs, with a regular spacing of about 5 km, are observed as well (Figure 7.21) (Nyffenegger and Consolmagno, 1988). Smooth material commonly occupies valley floors, in which case it exhibits a convex profile and can display medial ridges and/or troughs (Smith et al., 1986; Jankowski and Squyres, 1988).

A variety of models have been invoked to account for the ridges and troughs on Ariel. Both large- and small-scale ridges and troughs have been hypothesized to be the product of normal faulting (Smith et al., 1986; Nyffenegger and Consolmagno, 1988; Pappalardo, 1994). Some ridges within the ridged terrain might form by means of linear extrusion of viscous material (Ruzicka, 1988; Jankowski and Squyres, 1988). Smooth material occupying valley floors may have been emplaced as solid-state flows from linear vents (Jankowski and Squyres, 1988), perhaps
mobilized by interstitial volatiles (Stevenson and Lunine, 1986). Medial troughs on the smooth material may be due to faulting, perhaps related to the opening of fissures (Smith et al., 1986; Schenk, 1991), or they may be due to a “lava tube” style emplacement of smooth material (Croft and Soderblom, 1991). Isolated ridges in the smooth material may have formed as late stage extrusions (Smith et al., 1986; Schenk, 1991). Extensional tectonics and extrusion both imply a tensile stress state in the satellite’s lithosphere, perhaps induced by freezing of an initially molten interior (Smith et al., 1986; Plescia, 1987).

6.4 Dione, Tethys, Rhea, and Titania

We have grouped together Dione, Tethys, and Rhea (satellites of Saturn) and Titania (satellite of Uranus) because they all have relatively minor amounts of tectonism on their surfaces compared to the preceding bodies, and the origin of their tectonic features is still somewhat mysterious. All of them have surfaces dominated by impact craters, with isolated tectonic features cutting across the cratered terrain. On all of these satellites, the dominant tectonic features are organized sets of scarps (probably normal fault scarps) and graben.

From Voyager data, linear zones of bright lineations, termed “wispy terrain,” were mapped on Rhea’s and Dione’s surface (Smith et al., 1981; Plescia, 1983). In neither case were these features seen clearly by Voyager, but Cassini has revealed those on Dione to be sets of faults (Wagner et al., 2006; Moore and Schenk, 2007). The faults exhibit some bright scarps, and in some places are densely packed together in parallel groups (Figure 7.22). The main sets of faults on Dione appear to roughly follow a great circle, tilted with respect to the equator (Miller et al., 2007), although there are sets that clearly deviate from this pattern. Cassini images of Dione also reveal north–south trending ridges along the western boundary of a resurfaced plain that might be compressional (Moore and Schenk, 2007). Graben and normal fault sets on Rhea, along with several broad ridges identified in both Voyager and Cassini data (Moore et al., 1985; Moore and Schenk, 2007; Wagner et al., 2007) trend dominantly north–south.

On Tethys, the main tectonic feature is a large, wide graben complex called Ithaca Chasma (Figure 7.23). Cassini mapping shows that the graben system is roughly 2–3 km deep with a raised rim up to 6 km high (Giese et al., 2007; Moore and Schenk, 2007). Flexural modeling of the raised rim gives estimates of 16–20 km for the thickness of the elastic lithosphere during the formation of Ithaca Chasma (Giese et al., 2007). Ithaca Chasma is of interest because it is offset only 15–20° from a great circle centered on the large Odysseus impact basin (whose
Figure 7.22. Closely packed sets of normal faults on Saturn’s satellite Dione. Some faults are found in facing pairs, forming graben. North is to the top, and the illumination is from the right.

Figure 7.23. Ithaca Chasma on Saturn’s satellite Tethys is interpreted to be a large graben complex. North is to the top, and illumination is from the bottom right.
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Figure 7.24. Global view of the Uranian satellite Titania, showing several graben-like structures crossing its surface.

diameter is roughly 0.4 the satellite radius). This has led to suggestions that the chasma is tectonically related to the basin (e.g., Moore and Ahern, 1983; Moore et al., 2004a), though crater counts on the floor of Odysseus and the bottom of Ithaca Chasma indicate that the impact basin is younger (Giese et al., 2007).

Finally, Titania exhibits a branching network of faults and graben, 20–50 km wide and 2–5 km deep (Figure 7.24), which cut across most of the craters on the surface (Smith et al., 1986). The origin of these faults is unknown, due largely to the lack of a global image map of Titania.

These tectonic features, limited though they are, clearly imply something of the nature of the stress and thermal histories of these bodies. Freezing of water in the interior, the expansion of warming ice from radionuclide sources, been proposed for the expansion of these worlds’ interiors (Pollack and Consolmagno, 1984; Hillier and Squyres, 1991; Castillo et al., 2006). Reorientation of these bodies due to the mass asymmetries caused by large impact basins can produce stresses of ~100 kPa, which could potentially leave a record of surface fracturing in response, depending on the state of the interior at the time (Nimmo and Matsuyama, 2007). Ongoing mapping and analysis of these satellites will elucidate these issues and give us a better insight into the origin of tectonic features on middle-sized icy worlds.

7 Satellites without widespread tectonic activity

7.1 Titan

Titan is by far the largest satellite of Saturn, close to Ganymede in size, and is distinguished by its massive, extended nitrogen–methane atmosphere. Titan is an
active world, erasing craters from its surface at a geologically rapid pace. Only a handful of craters have been observed on the surface by the Cassini mission (Porco et al., 2005a; Stofan et al., 2006), though new ones are currently being discovered as our imaging coverage of the surface increases. Convincing evidence has been found for fluvial erosion features (Tomasko et al., 2005; Porco et al., 2005a; Stofan et al., 2006) and aeolian dunes (Stofan et al., 2006). Preliminary evidence has been found for cryovolcanic features (Stofan et al., 2006; Wall et al., 2009), but the evidence so far for tectonic activity on Titan is on shakier ground. Straight-sided features have been observed in near-infrared images (Porco et al., 2005a) but some of these have turned out to be fields of linear dunes (Lorenz et al., 2006). Eroded mountain chains have been observed in some of the radar data, but it is unclear what type of tectonic process formed them, or what the role of differential erosion is in determining the heights and shapes of these mountains (Radebaugh et al., 2007). Either Titan does not have active internally driven tectonic processes, or the erosional and depositional processes that modify Titan’s surface (e.g., aeolian, fluvial) are so effective at masking and erasing tectonic features that we cannot clearly determine the nature of Titan’s tectonics.

7.2 Callisto

Callisto, the outermost Galilean satellite, is similar in bulk properties to Ganymede, but its interior has not been differentiated as strongly into separate rock and ice layers. The dark, dusty surface is ancient and saturated with impact craters (see Moore et al., 2004b, for a full summary of Callisto geology). The most obvious tectonic features on Callisto are the concentric arcuate graben-like features that make up the multiring impact basins, such as Asgard and Valhalla (see subsection 3.3.4), somewhat similar to the furrows on Ganymede (subsection 6.1.1). Near the north pole, there is a group of narrow troughs, each several hundred kilometers long. They are oriented radial to a point on the surface, which may suggest an impact origin (Schenk, 1995), but the center of the system has not yet been imaged. If they are not impact related, they may be due to a late, slow, global expansion of Callisto, caused by a shutdown of internal convection (McKinnon, 2006). In this scenario, the colder poles would be favored locations for the tectonic expression of such late-stage expansion.

7.3 Mimas and Iapetus

Mimas and Iapetus are the innermost and outermost, respectively, of Saturn’s major moons. Both have heavily cratered surfaces, and both show hints of a minor episode of ancient tectonic activity. On Mimas, Voyager data revealed a global pattern of linear troughs across the surface that could either be related to tidal stresses or to
the large impact crater Herschel that dominates one side of Mimas (Moore et al., 2004a). Smith et al. (1981) suggested that large impacts could disrupt some of the inner satellites of Saturn. Large basins 400 to 600 km across are common on Iapetus and Rhea. Presumably, these impact events could induce incipient breakup fracturing on Iapetus, Rhea, and the smaller inner satellites, and Herschel on Mimas may be an example of this. Global mapping of recent Cassini images is ongoing to evaluate whether there is a link between the troughs and Herschel.

Other than scattered minor troughs and linear segments of crater walls, Iapetus displays only one major, possibly tectonic feature, but it is an impressive feature (Figure 7.25). A ridge up to 20 kilometers high runs exactly along the equator for more than one third of the satellite’s circumference (Porco et al., 2005b). Its mode of origin and the origin of the stresses that may have formed it are unknown and present an intriguing mystery (Ip, 2006). It may well be linked to the highly oblate shape of this moon (Castillo-Rogez et al., 2007) in a manner that is presently unclear.
7.4 Other satellites

There are plenty of other satellites in the outer solar system that have not been mentioned in this chapter, but most of them are small and irregularly shaped. The larger satellites we have neglected here, such as Umbriel and Oberon (satellites of Uranus), are very poorly covered by current imaging data, and so nothing definitive can be said at this point. Pluto and its major satellite Charon will not be visited by spacecraft until the New Horizons flyby in 2015. The orbital evolution of this “binary planet” may have induced significant tidal stresses on their surfaces (Collins and Pappalardo, 2000), and it will be interesting to see if there are tectonic features on the outer frontier of our solar system.

8 Conclusion

As we close this chapter, let us consider how tectonics on the outer planet satellites differs from tectonics elsewhere. Three overarching conclusions are (in no particular order):

- **Tides are supremely important**: There are many factors influencing the tectonics of the outer planet satellites that are governed by the giant planets around which they orbit. The stress fields that lead to the formation and global pattern of tectonic features on these satellites are often controlled or strongly influenced by changes in the tidal figure of the satellite. The evolution of a satellite’s orbit with time, which depends on dissipation within the giant planet and satellite and the orbital positions of a satellite’s siblings, is also very important, leading to changes in tidal heating and diurnal stresses. As these orbital and tidal parameters change over time, we cannot assume that any satellite has remained in a steady state with respect to tidal stresses and energy input. The immense amount of recent and ongoing geological activity in the outer planet satellites, even the tiny ones, is a testament to the power of giant planet tides to drive geological activity.

- **Tectonic features in ice are interpretable from terrestrial experience (mostly)**: Despite major differences in material properties, many of the tectonic features, such as isolated normal faults, folds, graben, and tilt-block complexes, are easily recognizable in the thick icy crusts of the outer satellites. However, there are some active satellites with thin elastic lithospheres (Europa, the south pole of Enceladus, possibly Triton) that exhibit bizarre tectonic features that defy easy comparison with terrestrial analogues.

- **Extension is ubiquitous, contraction is hard to find**: There is evidence on almost every outer planet satellite for some type of extensional tectonic feature, but only in a few places is there good evidence for strike-slip motion or surface contraction. The prevalence of extensional features may be due to the relative
ease of lithospheric failure in tension as opposed to compression, or perhaps there is a deeper message. For example, it is becoming increasingly apparent that subsurface oceans may be common on icy satellites. Freezing of such an ocean and thickening of the floating ice shell can cause tensile stresses near the surface (see subsection 3.2.6).

Much work remains to be done to unravel the tectonic history and current behavior of the outer planet satellites, and we have only just begun to explore this region of our solar system. For most satellites, we still lack complete global imaging coverage at a resolution sufficient to distinguish tectonic features. Various tools used for exploring terrestrial tectonics, such as close-up fieldwork, subsurface electromagnetic sounding, global gravity fields, and global altimetry data, have started to be used on extraterrestrial bodies, notably Mars, but none of the outer planet satellites have had such attention lavished on them yet. Current proposals for a new flagship mission focused on one of the outer planet satellites would go a long way toward filling the data gap. Obtaining a global imaging, radar sounding, gravity, and altimetry dataset for one of these outer planet satellites would give us a more solid foundation for understanding the behavior of icy lithospheres in general and the power of tides to shape them.

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References


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