15
Geology of Europa

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15.1 INTRODUCTION

15.1.1 Overview

Europa is a rocky object of radius 1565 km (slightly smaller than Earth’s moon) and has an outer shell of water composition estimated to be of order 100 km thick, the surface of which is frozen (Figure 15.1). The total volume of water is about $3 \times 10^9$ km$^3$, or twice the amount of water on Earth. Moreover, like its neighbor Io, Europa experiences internal heating generated from tidal flexing during its eccentric orbit around Jupiter. This raises the possibility that some of the water beneath the icy crust is liquid.

The proportion of rock to ice, the generation of internal heat, and the possibility of liquid water make Europa unique in the solar system. In this chapter, we outline the sources of data available for Europa (with a focus on the Galileo mission), review previous and ongoing research on its surface geology, discuss the astrobiological potential of Europa, and consider plans for future exploration.

15.1.2 Background

Europa has been viewed from Earth since its discovery through a simple telescope in the early 17th century by Galileo Galilei (Coyne 1997). Modern observations include those of Kuiper (1957) and Moroz (1965) who showed that Europa is probably covered with water frost. Reviews of these and more recent telescopic observations are given in relevant chapters in Burns (1977), Morrison (1982), Burns and Matthews (1986), and Barbieri et al. (1997).

The first close-up observations of the Jupiter system were from the Pioneer 10 spacecraft in December, 1973 (Fimmel et al. 1974), followed by Pioneer 11 one year later. These spacecraft flybys enabled new estimates of the masses and densities of the jovian moons and refinement of the concept that Europa is a rocky object with a surface of water ice. Although the flyby distances to Europa were too great to enable the Pioneer camera systems to obtain anything but cursory images, it was possible to discern a mottled dark and bright surface, indicating that materials other than water frost are present. Coupled with Earth-based determinations of Europa’s general properties, these new data led Cameron (1977) and Consolmagno and Lewis (1977) to derive models of Europa’s interior and its thermal evolution.

In 1979, Voyager 1 and 2 flew past Jupiter in March and July, respectively, and returned about 70 useful images of Europa. However, the best images (from Voyager 2) are only ~2 km/pixel (Smith et al. 1979a,b). Nonetheless, these were adequate to identify many of Europa’s unique characteristics, including the presence of lineaments longer than 1000 km and a marked paucity of large impact craters. Color data (McEwen 1986a) suggested the presence of non-ice materials on the surface. These and other geological characteristics are described by Lucchitta and Soderblom (1982) and
Malin and Pieri (1986). The relative lack of impact scars led to the idea that Europa’s surface is remarkably young, thought to be a result of extensive “resurfacing” by floods of liquid water erupted from the interior, a process referred to as cryovolcanism. Voyager results also enabled refinement of the interior models for Europa, as reviewed by Cassen et al. (1982), and Schubert et al. (1986), and consideration of the exobiological potential of Europa (Reynolds et al. 1983).

After a 6-year journey from Earth, the Galileo spacecraft began orbiting Jupiter in December 1995 and over the next 6 years made 18 flybys of Europa. This enabled the return of substantial remote sensing data, including images with scales as good as 6 m/pixel. Galileo image resolution and coverage are non-uniform, but most of the satellite has been imaged at better than 13 km per pixel, and most longitudes from 140° westward to 50° have been imaged at better than 1.6 km per pixel.

Although most of the geological considerations are based on the results from the Solid State Imaging (SSI) system (Belton et al. 1996, Carr et al. 1998) (see eTable 1 in accompanying CD) and the Near Infrared Mapping Spectrometer (NIMS) (Carlson et al. 1996), other instruments and data from the spacecraft also provide critical data. Particularly important for interpretations of the interior are data from the magnetometer (Kivelson et al. 1999) and the gravitational characteristics of the flybys from which the moment of inertia is determined (Anderson et al. 1998). Despite some significant hardware problems, the mission was incredibly successful (Chapter 1). Reviews of the general geology of Europa include Greeley (1997), Greeley et al. (1998a, 2000), and Pappalardo et al. (1999a,b). The results from the Galileo mission include proposed identifications of non-ice compositions on the surface, including various salts (e.g., McCord et al. 1999), the discovery of numerous surface features, only some of which can be explained readily, and elucidation of the interior and its evolution (Anderson et al. 1998). Particularly noteworthy is the debate concerning the presence or absence of liquid water beneath the ice crust. As reviewed previously by Pappalardo et al. (1999a,b), while the evidence for liquid water in the past is favorable, there is no unambiguous indication from spacecraft imaging that such conditions exist today. Interest in Europa for future exploration is high because the satellite could contain environments conducive for prebiotic and exobiological evolution. These and other issues are discussed in the following sections.

### 15.1.3 Sources of Spacecraft Data

Most of the information relevant for the geology and geophysics of Europa is derived from the Voyager and Galileo missions as archived in the Planetary Data System (McMahhon 1996). On the accompanying CD, eFigure 1 list and show the SSI image foot prints for Europa from the Galileo mission, while eFigure 2 shows the Galileo NIMS data. eTable 2 lists the Voyager images obtained for Europa.

### 15.2 SURFACE PROPERTIES

The surface of an object is a window to its interior. For example, evidence of the interior evolution is seen in the type and distribution of surface features, composition, and mineralogy. The surface is also an interface between the interior and the atmosphere and the space above it. Consequently, it is a rich zone for chemical and physical processes, which are often sensitive to temperature, pressure, and other environmental factors. Although considerable information about the bulk surface properties of Europa has been determined over the last 50 years using Earth-based telescopes, the knowledge has been vastly expanded recently by the Voyager and Galileo missions (cf. McCord 2000). Earth orbital telescopes have also made valuable contributions recently, even with the limited spatial resolution possible.

Before spacecraft visited Europa, the surface was known from ground-based spectrophotometry to be mostly water ice (Kuiper 1957, Harris 1961, Moroz, 1965, Johnson and McCord 1971, Pilcher et al. 1972, see reviews by Morrison and Morrison 1977, Calvin et al. 1995, McCord et al. 1998a, 1999, McCord 2000). An unidentified non-water ice component was suspected, especially on the trailing side, but was reported to be present in minor amounts (e.g., Clark 1980).

The daytime surface temperature was found to range from
surface temperatures, but the data were noisy and consisted
of the surface of Europa contains considerable amounts of a
thermal inertia related to thermal conductivity, perhaps
equator. These are most likely due to variations in
the International Ultraviolet Explorer (IUE) and the Hub­
systematic variations with latitude, with higher post-sunset
showed low-latitude diurnal brightness temperatures in the
son and Brown (1989) proposed a solid-state greenhouse
explain the eclipse observations (Tamppari et al. 1995).
The Voyager mission returned some additional data on
surface temperatures, but the data were noisy and consisted
of hard-to-interpret non-blackbody spectra (Spencer 1987a).

Many images from the Voyager cameras showed com­
plex surface features, including geologic/compositional units
with sharp boundaries, but little information on specific
surface composition. The Galileo NIMS has provided the
most detailed and extensive information about the surface
composition so far available, augmented by the Ultraviolet
Spectrometer (UVS) and Earth-orbital spectroscopy using
the International Ultraviolet Explorer (IUE) and the Hub­
ble Space Telescope (HST). They have generally confirmed
the earlier findings, except that there appear to be more
extensive non-water-ice materials than originally thought.

The Galileo Photopolarimeter-Radiometer (PPR)
showed low-latitude diurnal brightness temperatures in the
range of 86 to 132 K (Figure 15.2) (Spencer et al. 1999).
These are thought to be near the kinetic temperatures
because of the high emissivity (\( \geq 0.9 \)) of water ice in the IR.
Daytime temperatures are inversely correlated with surface
albedo, as expected. On a global scale, this is true for
nighttime temperatures as well. Significant local variations
in temperature below the spatial resolution of the PPR
are likely due to local distributions of darker and lighter
materials (Spencer 1987b). There are deviations at night
from the darker-is-hotter trend, such as the bright but warm
jecta blanket near the crater Pwyll, perhaps due to crater
jecta and surface boulders. Most difficult to explain are
systematic variations with latitude, with higher post-sunset
temperatures at northern latitudes than at corresponding
southern latitudes and with a temperature minimum along
the equator. These are most likely due to variations in
thermal inertia related to thermal conductivity, perhaps
due to grain-to-grain contact variations.

In addition to water ice with a range of particle sizes,
the surface of Europa contains considerable amounts of a
hydrated mineral(s). This is evident from the highly-dis­
Figure 15.2. Contours of brightness-temperature distributions
on Europa obtained from the Galileo Photopolarimeter-
Radiometer (from Spencer et al. 1999). At the time of going to
press a colour version of this figure was available for download
from http://www.cambridge.org/9780521035453.

torted water absorption bands present in the NIMS spec­
tra (Figure 15.3). McCord et al. (1998b, 1999) reported on
the nature and distribution of these spectral features and
interpreted them to indicate the presence of hydrated salt
minerals, sulfates, and perhaps carbonates. They and Fanale
et al. (2000) showed these hydrated deposits to be associ­
closedly with the lineaments and mottled/chaos terrain,
regions of most recent disruption (Figure 15.4). Thus, the
hydrates were interpreted to be due to endogenic processes,
probably related to a briny ocean below the surface. How­
ever, it was suggested by Carlson et al. (1999b) that sulfuric
acid (\( H_2SO_4 \)) hydrate, created by radiolysis of sulfur from
I0, processing of endogenic \( SO_2 \), or from the previously sug­
gested sulfate salts, also or instead could be present in these
deposits (Chapter 20).

Several investigators expressed concern that hydrated
salt minerals might readily undergo dehybridation and/or
disassociation on Europa’s surface. McCord et al. (2001b)
showed experimentally that at least some of the hydrated salts,
especially doubly-bonded materials such as magnesium sulfate
hydrate, are more stable than water ice to dehydration and
could exist over the age of the solar system on the surface of
Europa. This result was in agreement with theoretical cal­
culations (Zolotov and Shock 2001). Thus, lag deposits with
increasing concentrations over time might be expected and
could explain the apparently high concentrations of salts in

Figure 15.3. Reflectance spectra for the Europa hydrated min­
eral deposits compared with those for a variety of water ice grain
sizes (from McCord et al. 1999).
some areas. Moreover, it was found that magnesium sulfate brines rapidly quenched on cold surfaces under vacuum produce poorly-crystalline and heavily hydrated materials with reflectance spectra even more like those for Europa than for crystalline samples produced at room temperature (McCord et al. 2002).

Magnesium sulfate hydrate is expected in large quantities from the thermal evolution and aqueous differentiation of an Europa-like body composed initially of carbonaceous chondrite material (Fanale et al. 1977, 2001, Kargel 1991, 1999, Kargel et al. 2000, Zolotov and Shock 2001, McKinnon 2002). From spectral evidence, geochemical modeling, meteorite leaching experiments (Fanale et al. 1998) and geological associations, McCord et al. (2001b) postulated that a mixture of mostly MgSO₄·nH₂O, with some Na₂SO₄·nH₂O, is the most likely hydrate present on Europa. Nevertheless, from the reflectance spectrum alone, it is not possible to rule out sulfuric acid and perhaps other materials being present. In fact, if Na₂SO₄ were present, some singly bonded Na is likely to be radiolyzed, H⁺ taking its place, producing sulfuric acid (see Chapter 20). Thus, both salt minerals (mostly MgSO₄) and some sulfuric acid (H₂SO₄), derived from Na₂SO₄·nH₂O, could be present as hydrates. The hydrated salt minerals seem to exhibit a grain size in the 300-500 μm range (Van Keulen et al. 2000).

Various minor molecular constituents have also been discovered in the surface materials of all three icy Galilean satellites. They are apparently both indigenous and due to radiation processing. CO₂ has been reported (Carlson et al. 1996, McCord et al. 1997, 1998b, Smythe et al. 1998, Hibbits et al. 2000) and has been found escaping from at least Callisto (Carlson 1999). Other constituents (SO₂, S-H, C≡N, C-H) were also reported for Ganymede and Callisto from absorptions in the 3-5 μm region (McCord et al. 1997, 1998a), but the near-complete coverage of Europa’s surface by H₂O-bearing minerals, which absorb strongly in this spectral region, allows little radiation to be reflected, and thus signatures of minority materials are easily hidden. Nevertheless, Carlson et al. (1999a) reported H₂O₂ from a 3.50-μm absorption visible in regions containing fine-grained water ice (approximately the leading side). The UVS detected spectral evidence of H₂O₂ on all three icy satellites (Hendrix et al. 1999), which is probably a product of radiolysis of water ice (Moore and Hudson 2000). The UVS team also reported a UV absorber concentrated on the trailing side of Europa (Hendrix et al. 1998), consistent with the UV-absorbing surface stain on the trailing side mapped using the Voyager multispectral images (Johnson et al. 1983, Johnson et al. 1988, McEwen 1986a, Nelson et al. 1986) and attributed to sulfur delivered to the surface from Io by the Jupiter magnetosphere.

The processes affecting Europa’s surface materials are generally related to thermal processing and radiation bombardment. Over a longer timescale, meteorite impact would be expected to influence the nature of the surface, but the rapid resurfacing of Europa, probably due to interaction with a warm ocean below, might dominate. The temperature range for Europa’s surface (~90 to 130 K) is in the critical range in which the properties of water ice and the chemistry induced by low-energy electrons change rapidly with temperature. Thus, the surface processes are likely to be complex and interactive, making Europa a rich laboratory for studying material properties.

Perhaps the most important environmental factor is the intense, high-energy (up to 10 MeV) radiation flux impacting the surface due to trapped particles in Jupiter’s magnetosphere (Cooper et al. 2001). This includes electrons and protons as well as heavy ions such as O and S. A major process, especially due to the heavy ions, is sputtering, i.e., physical disruption of the surface molecules tending to blast fragments from the surface (Chapter 20). One consequence is the creation of an exosphere of sputtered products such as Na (Brown and Hill 1996, Johnson 2000), possibly from Na₂SO₄ hydrate, that could be used as a probe of the surface constituents using mass spectrometers on or near the surface (Johnson et al. 1998).

Low-energy electrons are also produced as secondaries by sputtering as well as induced directly. These are the most important for causing reactions that change the chemical (rather than the physical) nature of the materials. Extensive research on this surface chemistry is developing, and is elaborated on in Chapter 20. For example, Moore and Hudson (2000) and studied H⁺ irradiation of ices to produce H₂O₂ and McCord et al. (2001b) and Zolotov and Shock (2001) investigated the chemistry and stability of hydrated salt minerals under European conditions.

The physical nature of the surface is important for studies of thermal and optical radiation transport and other processes. Water ice on the surface of Europa exhibits a variety of grain sizes. The ice between ±60° latitude is generally abundant and fine-grained (radius <50 μm) on the leading side (Hansen et al. 1998, Shirley and Carlson 2000) but less abundant, well segregated, and large grained (radius >200 μm) on the trailing side (e.g., McCord et al. 1999). The polar caps have an unusual water-ice spectrum, which is likely the result of mixtures or layering of frost particles with a large range of grain sizes. The top micrometer of surface ice is amorphous over all of Europa (Hansen and McCord 2000). This is probably an indication of the relatively high radiolytic disruption of the ice crystalline structure compared to the other icy satellites, where there is much more
crystalline surface ice. Thermal annealing and recrystallization of the ice can counteract the radiolytic disruption, but this is less operative on Europa because of its cold temperatures.

Water molecules from water ice and to a lesser extent from hydrated minerals are removed by sputtering and by thermal desorption, but probably mostly re-condense elsewhere. The redistribution of water molecules probably creates an underdense, frosty surface that slowly sinters with time. This effect will vary with temperature, i.e., with latitude and albedo, producing polar frost deposits for example. Further, it appears that the dark lineaments brighten with age (Greeley et al. 1998b, Geissler et al. 1998a), which could be due to water frost slowly covering the older surfaces. One consequence might be to produce free H2O molecules that create a "humid" layer just above the surface and induce re-hydration, slowing the overall dehydration process.

Therefore, study of Europa's surface has revealed a rich set of processes that are important for the other icy Galilean satellites as well as for other cold surfaces in the solar system. In comparison with Europa, Ganymede's surface (see Chapter 16) seems to be mostly water ice but with some hydrated minerals, probably salts, also present in the darker regions (McCord et al. 2001a). Ganymede has prominent polar frost caps, possibly indicating transport of water as a vapor. Callisto (see Chapter 17) is much less icy, with over 50% of the surface covered with dark material having a strong O–H absorption near 3 μm but no detectable H2O features, suggesting hydroxide minerals such as clays. Both Ganymede and Callisto are reported to have various minority constituents, including H2O2, CO2, SO2, CH, CN, and oxygen (see summary in McCord 2000); these could be on Europa as well, but their detection is difficult.

15.3 GEOLOGY
15.3.1 Surface Features

Global Perspective

The highest resolution Voyager images of Europa revealed a variety of colors (McEwen 1986a) and morphological features. Lucchitta and Soderblom (1982) divided the surface into plains (generally bright areas transected by linear features) and mottled terrain (darker speckled units which generally seemed to be superimposed on older plains). Lineaments were subdivided by their albedo, planform, and albedo-based classifications to be disentangled in the region of their overlap, as the troughs taper in width and intersect, presumably reflecting interaction of the stress fields of simultaneously propagating tension fractures. The elevated flanks of some troughs appear to be transitional to the formation of some double ridges (Head et al. 1999, Geissler et al. 1998a).

Troughs

The simplest of Europa's landforms are linear to curvilinear troughs (Figure 15.6). They are generally V-shaped in cross section, 100–300 m wide, and can be >100 km long. Their rims are level or raised with respect to the surrounding terrain. These characteristics suggest an origin as tension fractures. The width of most troughs implies that they became wider after the initial fracturing, perhaps from mass wasting and/or movement along the trough axis. In some areas, en echelon troughs curve inward toward one another in the region of their overlap, as the troughs taper in width and intersect, presumably reflecting interaction of the stress fields of simultaneously propagating tension fractures. The elevated flanks of some troughs appear to be transitional to the formation of some double ridges (Head et al. 1999, Geissler et al. 1998a).

Ridges

Ridges are Europa's most ubiquitous landform (Figure 15.6) but are not well understood. They most commonly take...
Figure 15.5. (a) Polar and (b) mercator maps of Europa showing key place names (from U.S. Geological Survey).
the form of double ridges, i.e., a ridge pair with a medial trough (Figure 15.6a). Some classifications (Pappalardo et al. 1998b, Greenberg et al. 1998) suggest an evolutionary transition from isolated troughs, to double ridges (including "triple" bands), to complex ridges which generally show a series of subparallel features. Ridges can be short or can have lengths of >1000 km. Their great along-trend uniformity poses a challenge to ridge formation models.

Double ridges are ~500 m to ~2 km wide, have flank slopes near the angle of repose for loose, blocky material, and are characterized by a continuous axial trough. In cross section, some double ridges are slightly convex to trapezoidal, with a central depression. Pre-existing topography has been identified on some ridge flanks (Head et al. 1999), though this interpretation has been disputed (Sullivan et al. 1999). Mass wasting is prevalent along the ridges, with the debris apparently draping over pre-existing terrain (Sullivan et al. 1999). In some cases the trend of the pre-existing topography appears to be deflected as the ridge flank is encountered (Figure 15.6b, full arrow). Many ridges show evidence for strike-slip motion (e.g., Figure 15.6b, half arrows), a characteristic not seen in isolated troughs (Hoppa et al. 1999a). Extension is clearly indicated by pre-existing structures offset by some wide ridges that show complex, lineated morphology (Pappalardo et al. 1998b, Tufts et al. 2000).

Apparent ridge morphology is sensitive to lighting geometry. At low solar incidence angle, some complex ridges appear as triple bands (Figure 15.6c) with diffuse dark material flanking the ridge and infilling topographic lows (Greenley 1997). In high resolution stereo images, it is apparent that dark material is situated in local topographic lows, such as the floors of axial troughs and on wall terraces (Figure 15.6d). The ridge of Figure 15.6e appears to change morphology where it changes direction at a cusp; the apparent change could be due to the orientation with respect to the illumination. The cycloidal pattern of some ridges (e.g., Figure 15.6e) suggests they originated as cycloidal fractures influenced by Europa’s diurnally rotating stress field (see Section 15.3.4), while straight ridges probably evolved from relatively straight cracks, perhaps formed with initially greater propagation velocity (Hoppa et al. 1999b). Some ridges are flanked by topographic depressions and/or fine-scale fractures (Figure 15.6f), suggesting loading of the lithosphere either from above (due to the weight of the ridge material) or from below (e.g., due to withdrawal of subsurface material) during ridge formation.

We emphasize that cracking patterns and ridge formation could be related, but are inherently different issues. As discussed below, quantitative investigations of stress models and implied cracking patterns indicate that Europa’s lineaments probably originated as tensile or extensional tectonic features formed and opened by diurnal and nonsynchronous stresses (Helfenstein and Parmentier 1980, 1983, 1985, McEwen 1986b, Leith and McKinnon 1996, Greenberg et al. 1998, 2002a, Hoppa et al. 1999a,b). The inference that ridges have evolved from isolated troughs suggests that ridge patterns reflect the major crack patterns within Europa’s ice shell. Cracking models are broadly consistent with a range of possible ridge formation models and do not demonstrate any specific ridge formation model. Successful formation models must take into account the observed ridge morphologies, inferred transitions among ridges and other structures, geologic setting and associations, and geophysical constraints.

At least six models have been proposed for the origin of Europa’s double ridges (a topic reviewed by Pappalardo et al. 1999a). Each model has different 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, and some imply that liquid water exists in the shallow subsurface.

**Tidal squeezing.** Greenberg et al. (1998) suggest that fractures penetrate through an ice shell and open and close by ~1 m in response to diurnal stresses, allowing water and icy to be pumped toward the surface with each cycle to build ridges. Because this model explicitly assumes that the ice shell of Europa is penetrated by dry cracks, the model necessitates a thin ice shell, because cracks can penetrate at most ~6 to 10 km (Golombek and Banerdt 1990, Leith and McKinnon 1996). Crawford and Stevenson (1988) demonstrated the difficulty of cracks penetrating the ice shell, notably in cracking through warm ductile ice near the ocean interface. They found that liquid-filled cracks could ascend to the surface, but the cracks cannot be held open through the entire ice shell. Other difficulties with the model are that diurnal tidal stresses appear insufficient to crack the shell to more than ~150 m depth – although stresses resulting from nonsynchronous rotation may open surface tension cracks to significantly greater depth (see Section 15.4.2), and water in narrow cracks is expected to freeze faster than the tidal cycle (Gaidos and Nimmo 2000).

**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 CO₂ or SO₂ are capable of driving eruptions, despite the negative buoyancy of water relative to ice (Fagents et al. 2000). Like the tidal pumping model, the volcanic model has difficulty in cracking the ice if it is presumed that open conduits extend from a subsurface ocean to the surface. It is conceivable 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 has difficulty accounting for the great length of some ridge complexes across Europa’s surface.

**Dike intrusion.** Turtle et al. (1998) consider ridges to form by intrusion of melt into a shallow vertical crack to build a double ridge. In this model, a subsurface dike intrudes a shallow subsurface crack that is constricted at its bottom, causing outward and upward plastic deformation to build a ridge.

**Compression.** Sullivan et al. (1998) propose that ridges are compressional structures. Compression is a viable model for some ridges, based on reconstruction of pre-existing features (Patterson and Pappalardo 2002) and kinematic arguments (Sarid et al. 2002).

**Linear diapirism.** Head et al. (1999) propose that double ridges form in response to cracking and diapirc rise of warm ice, which intrudes and lifts the surface to form ridges. This model suggests that cracks penetrate to a subsurface ductile ice layer, rather than through the ice shell. Warm subsurface ice moves buoyantly into the fracture, aided by tidal heating. The process is envisioned as anal-
Figure 15.6. Examples of ridge and trough morphologies and characteristics seen on Galileo images: (a) typical double ridges observed near the terminator (11ESMORPHY01; 33 m/pixel); (b) a complex ridge with possible uplift of pre-existing ridge structures along its flanks (full arrow), and a ridge showing evidence for strike-slip displacement (half arrows) of pre-existing structures (17ESSOUTHPO1; 40 m/pixel); (c) a triple band seen at high resolution and low solar incidence angle (14ESTRPBND01; 70 m/pixel); (d) a double ridge seen at low solar incidence angle, revealing dark mass wasted material (arrow) within its crestal trough (12ESWEDGE-02; 25 m/pixel); (e) the cusp of a cycloidal ridge that crosses Astypalaia Linea, with arrows indicating the ridge axis where topography is less prominent (17ESSTRSLP01; 40 m/pixel); (f) a ridge with flanking cuspate troughs and associated fractures (arrows), suggesting flexural downwarp of the brittle surface layer (E6ESBRTPLN01; 53 m/pixel). Some images have been rotated with respect to north, so that illumination is from the left.

ogous to the rise of tabular-shaped “salt walls” which rise along extensional fractures on Earth (e.g., Jenyon 1986) and can cause intrusive uplift. This model has been criticized on the grounds that inner ridges would not be expected of upwelling diapirs (Greenberg et al. 2002a).

Shear heating. Gaidos and Nimmo (2000) and Nimmo and Gaidos (2002) modify the diapirism model in suggesting that diurnal motion along tidally deformed fractures induces shear heating sufficient to trigger upwelling of warm ice to form a ridge, and perhaps induce partial melting. Frictional heating along a discrete fracture will reduce the ice viscosity within a corresponding shear zone of finite width. The warm,
low-viscosity ice can rise buoyantly to create a ridge, potentially building a few-hundred-meter high ridge in 10 years. Shear heating could enable the intrusion, uplift, and extrusion proposed in the linear diapirism model (Pappalardo et al. 1998b, Head et al. 1999).

**Triple bands.** Galileo observations provide insight into the nature and origin of Europa's triple bands. Their dark margins are seen to be diffuse, and their specific morphologies diverse (Greeley et al. 1997, 1998a,b). Although a ridge load might depress a thin ice lithosphere below a water line to cause surface flooding (Greenberg et al. 1998), Pappalardo and Coon (1996) argue that this is possible only if the lithosphere is \( \leq 2 \) km thick. Fagents et al. (2000) considered the possibility that the dark flanks of triple bands were created by ballistic emplacement of dark materials entrained in gas-driven cryovolcanic eruptions, or that they are thin dark lag deposits formed adjacent to a water or solid-state ice intrusion due to sublimation of surface frosts and local concentration of refractory materials. Along the lines of a model suggested by Head and Pappalardo (1999), ridge-related heating might trigger partial melt generation, perhaps mobilizing brines that contribute to formation of dark flanks.

**Bands**

The geometries of dark wedge-shaped and gray bands on Europa enable reconstruction of their original configuration, restoring structures that were apparently displaced as the bands opened along fractures (Schenk and McKinnon 1989, Pappalardo and Sullivan 1996, Sullivan et al. 1998). These features have been termed “pull-aparts” (Greeley et al. 1998a,b) or simply “bands” (Prockter et al. 2000). The morphological relationships of pull-apart bands are most clearly seen southwest of Europa's antijovian point, where small plates separate darker material (Schenk and McKinnon 1989, Belton et al. 1996, Sullivan et al. 1998, Tufts et al. 2000). Reconstructions of pull-apart bands imply that the surface layer has behaved brittlely, separating and translating ice slabs underlain by a low-viscosity subsurface material, with the region of separation being infilled with relatively dark, mobile material. Thus, pull-apart bands offer compelling evidence for warm, mobile material in the shallow subsurface at the time of their formation. Stratigraphic evidence indicates that dark bands brighten with time (Pappalardo and Sullivan 1996, Greeley et al. 1998a,b, Geissler et al. 1998b), perhaps related to chemical changes and/or sputtering processes that redistribute surface frost (Geissler et al. 1998a, Pappalardo et al. 1999a).

Analyses have been made between pull-apart formation on Europa and the formation of leads in terrestrial sea ice (Pappalardo and Coon 1996, Greeley et al. 1998a,b), 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. The cracks are unable to close completely during the compressional phase due to the addition of new material; hence, the band widens with time as new material is added. However, as described above, models which rely on cracking through the ice shell to the depth of liquid water face a number of objections.

Arguing from morphological analyses of Galileo images, Sullivan et al. (1998) and Prockter et al. (2002) suggest that band formation could be analogous to terrestrial spreading centers. In terrestrial plate tectonics, “spreading” occurs where the lithosphere has separated and new material is intruded and extruded as the opposing original lithospheric plates are pulled apart. Regional Galileo images of pull-apart bands show an internal structure of ridges and troughs trending subparallel to each other and the boundaries of the band (Figure 15.7a). An overall bilateral symmetry suggests that the spreading analog may be appropriate in some cases.

Pull-apart band margins are generally sharp, and some show rounded bounding ridges (Figures 15.7a,c). A narrow central trough is common and is remarkably linear and uniform in width (Figures 15.7b,c,e, arrows). A hummocky textured zone commonly occurs on either side of the trough (branches on Figures 15.7b,c), or sometimes throughout much of a band (Figure 15.7d). Outside of the hummocky zone (Figures 15.7b,c,d) are subparallel ridges and troughs 300 to 400 m wide, which are regularly spaced and generally triangular in cross section. In the band of Figure 15.7d, the ridges and troughs have been interpreted as south-facing domino-styled fault blocks (Figueroa and Greeley 2000, Prockter et al. 2000). Some bands, notably Astypalaea Linea (Figure 15.7e), show a significant strike-slip component, suggesting oblique opening (Tufts et al. 1999).

Reconstruction of bands suggests that at least some have opened along pre-existing ridges (Prockter et al. 2002). Galileo stereo images of several dark pull-apart bands (Figures 15.7a and 15.7b) show that bands commonly stand higher than the surrounding ridged plains (Giese et al. 1999, Prockter et al. 2000, Tufts et al. 2000). This is consistent with emplacement of buoyant material, such as ice that is warm and/or clean relative to the cold and/or saltier surrounding lithospheric material (Prockter et al. 2000). These bands can cause significant tectono-volcanic resurfacing of Europa (Pappalardo and Sullivan, 1996, Greeley et al. 2000). The bright ridged plains of Europa contain many bands and band fragments; however, ridge morphologies in the ridged plains are diverse, suggesting a complex origin through multiple processes.

Agenor Linea is a rare bright band on Europa's surface, considered from Voyager analyses to be possibly of compressional origin (Schenk and McKinnon 1989). It has unusual photometric properties suggesting that it might have been active recently (Geissler et al. 1998c). However, high-resolution observations during Galileo orbit E17 show that it is cross-cut by small fractures and mottled terrain, reducing the likelihood that it is currently active (Prockter et al. 1999). Structural relationships within the band suggest that it formed by right-lateral strike-slip motion of ice slabs (Prockter et al. 2000), whereas Sarid et al. (2002) present evidence for a component of convergence.

**Folds**

Extensional tectonics was identified on Europa beginning with the Voyager mission (e.g., Schenk and McKinnon 1989). However, only tentative examples of compressional structures have been identified. Morphological evidence for folds with \(~25\) km wavelength is found in the Astypalaea Linea region (Prockter and Pappalardo 2000) (Figure 15.8).
Figure 15.7. Examples of band morphologies and characteristics seen on *Galileo* images: (a) a wedge-shaped dark band (DB) cross-cuts a brighter gray band (GB); outlines mark locations of the two subsequent views (C3ESWEDGES01; 420 m/pixel); (b) at high resolution, the dark band shows multiple subparallel ridges and troughs that flank a central trough (arrows) and hummocky zone (braces, HZ) (12ESWEDGE-02; 25 m/pixel); (c) at high resolution the gray band shows similar characteristics of a central trough (arrow) and hummocky zone (braces, HZ) (12ESWEDGE-02; 25 m/pixel); (d) a band seen at high resolution and low solar incidence angle displays subparallel ridges and troughs with characteristics of normal fault blocks (11ESMORPHY01; 33 m/pixel); (e) Astypalaea Linea has probably opened by strike-slip motion (arrows) along a pre-existing double ridge (17ESSTRSLP01; 40 m/pixel).
Europa

Figure 15.8. Possible folds on Europa: (a) subtle shading variations across Astypalaia Linea are inferred to be folds of ~25 km wavelength (T7ESSTRSLP01; 40 m/pixel); (b) a low-pass filter highlights the long-wavelength anticlinal swells and synclinal valleys (after Prockter and Pappalardo 2000).

Galileo images show subtle shading variations suggestive of folds, including anticlines and synclines. Relaxation of fold topography could cycle material back into Europa’s deeper interior, of relevance to astrobiological models (see Section 15.5).

Other folds could exist (Figueredo and Greeley 2000, Prockter et al. 2002) but subtle long-wavelength topography makes them generally difficult to recognize. Some rounded ridges in the plains could be small folds. Overall, these folds can accommodate only small amounts of strain. Possible convergent bands have also been identified by Greenberg et al. (2002b) and Sarid et al. (2002), and may help alleviate the overall strain problem.

Pits, Domes, and Spots

Galileo images show that Europa’s domes, pits, and spots (commonly referred to as “lenticulae”), along with larger chaos regions and smooth dark plains, comprise the satellite’s mottled terrain (Figure 15.9, Carr et al. 1998, Pappalardo et al. 1998a, Greeley et al. 2000). Pappalardo et al. (1998a) argue that most lenticulae have diameters of ~10 km (see also Spaun et al. 1999a,b, 2001), and that this size similarity and morphological gradation among pits, domes, and spots suggests that they are genetically related (Carr et al. 1998, Greeley et al. 1998b, Pappalardo et al. 1998a), which would be consistent with an origin through diapirism as the manifestation of solid-state convection of Europa’s icy shell (Rathbun et al. 1998). However, Greenberg et al. (1999) argue that pits and spots are only small members of a continuous size distribution of chaos areas.

The domes have been subdivided into two endmember types based on their relationship to the pre-existing surface (Carr et al. 1998). Type 1 domes consist of material different from the ridged plains. Typically this material is darker than the ridged plains. Pre-existing terrain has been destroyed in situ and/or obscured by newer material. Proposed models of formation include: (1) ice volcanism (Fagents et al. 2000), (2) diapirism (Pappalardo et al. 1998a, Rathbun et al. 1998) possibly accompanied by partial melting of a salt-rich icy lithosphere (Head and Pappalardo 1999), and (3) melt-through of the ice shell (Carr et al. 1998, Greenberg et al. 1999). The claim (Pappalardo et al. 1998a) for regular size and spacing suggests diapiric origin, a mechanism predicted by models of tidally heated ice above liquid water (Reynolds and Cassen 1979, Squyres and Croft 1986,
Figure 15.9. *Galileo* views of lenticulae: (a) regional view of an area dense with lenticulae, illustrating their common \(\sim 10\) km diameters and morphological variety as pits, domes, and spots (15ESREGMAP01; 230 m/pixel). (b) a dome showing evidence for localized uplift of the pre-existing ridged plains; (c) pits showing surface disruption and/or extrusion to create hummocky material; (d) lenticula with an annulus of smooth, relatively low albedo material, suggesting thermal alteration of the surface or localized extrusion of melt; (e) lenticula with convex margins that stand slightly above the surrounding terrain, suggesting extrusion of viscous material; (f) high-resolution view of a lenticula illustrating in situ disruption of pre-existing ridged plains and replacement by hummocky material, in a manner similar to the formation of larger chaos regions (11ESMORPHY01; 33 m/pixel). Illumination is from the right in all images; north is toward the top in all images except (f), which is south toward the top.


Pre-existing terrain is preserved on the surfaces of many domes, suggesting that the surface was upwarped but not destroyed (Carr *et al.* 1998, Greeley *et al.* 1998b, Pappalardo *et al.* 1998a, Pappalardo 2000). These Type 2 (Carr *et al.* 1998) or "upwarped" (Pappalardo 2000) domes are subcircular to elliptical, locally high features on which continuous pre-existing structure is preserved. Fractures can occur along their crests. Dome boundaries can be continuous with respect to the surrounding terrain with no discrete scarp, producing a gentle convex dome, or a relatively flat-topped surface. These morphologies suggest upwarp and flexure of an elastic surface layer. Domes are commonly bounded by abrupt scarps of apparent tectonic origin. Associated dome tops are tilted or relatively flat, suggesting brittle failure and punching upward of the dome along bounding fractures, consistent with laccolith-like intrusions (Pappalardo 2000).

Chaos regions are typically composed of polygonal blocks of pre-existing ridged plains that have shifted within a matrix of hummocky material (Carr *et al.* 1998). Chaos matrix material can be either low-lying or high-standing relative to the surrounding plains (Collins *et al.* 2000, Greeley *et al.* 2000). In Conamara Chaos (Figure 15.10), \(\sim 60\%\) of the pre-existing terrain was replaced or converted into matrix material, and many of the surviving blocks can be restored to their original positions (Spaun *et al.* 1998).

Morphological transition to larger chaos areas suggests related formational processes for lenticulae and chaos (Spaun *et al.* 1999a, Riley *et al.* 2000). Chaos regions have been interpreted as areas of amplified heat flow and perhaps local melting (Carr *et al.* 1998, Williams and Greeley 1998, Greenberg *et al.* 1999). In a melting model, blocks are analogous to buoyant icebergs. Alternately, solid-state ice
might have risen diapirically to the surface, disrupting the relatively cold and rigid lithosphere similar to the means proposed for lenticulae (Pappalardo et al. 1998a) and translating fragmented slabs of colder lithospheric material. The conversion of pre-existing ridged terrain to chaos appears to involve separate disruptions that merge (Spaun et al. 1999a).

Complete melt-through of Europa's ice shell would require a weakly stratified ocean to permit transfer of heat from the silicate mantle to the base of the ice shell, and a large plume of concentrated heat that is stable for at least hundreds of years (Collins et al. 2000, Thompson and Delaney 2001, O'Brien et al. 2002, Buck et al. 2002). If the ice shell is thick, flow of the warm base of the ice shell would be so rapid as to preclude melting (Stevenson 2000b). However, this mechanism does not preclude melt-through in a thin <6 km ice shell (O'Brien et al. 2002).

The fact that small blocks were mobilized and tilted within the chaos challenges the notion that the matrix was emplaced as solid-state ice, because the timescale of block movement in warm ductile ice is expected to be longer than the timescale of thermal diffusion causing blocks to cool in place (Collins et al. 2000). In models without melt-through, it is more likely that chaos formed in response to local melting within the ice shell, triggered by rising warm diapiric masses. This could be partial melting as the diapir impinges on relatively salt-rich ice (Head and Pappalardo 1999, Collins et al. 2000). Melting due to concentrated tidal heating within warm ice diapirs was suggested (McKinnon 1999, Wang and Stevenson 2000, Sotin et al. 2002), but the small size of diapirs could mean that their tidal heating is negligible (Moore 2001).

Topographic data across Conamara (Figure 15.10d) provide evidence that much of the chaos is ~300 m higher than the surrounding plains (Schenk and Pappalardo 2002). Similar results are found for a mitten-shaped chaos region in the leading hemisphere (Figueroedo et al. 2002). High-standing topography is hard to understand if chaos formed by melt through of an ice shell and then allowed to cool to thermal steady state, but can occur if warm diapiric ice rises from the base of ice shell and is extruded.

15.3.2 Age Relationships

The existence, stability, and evolution of a subsurface ocean through time are intricately tied to Europa's surface age and geologic activity. Most chaotic and dark plains materials are among Europa's youngest units, cross-cutting older bands and ridged plains, while ridged plains materials are commonly inferred to be the oldest units (Figure 15.11) (Head et al. 1999, Sullivan et al. 1999, Prockter et al. 2000, Figueroedo and Greeley 2000, Kadel et al. 2000, Greeley et al. 2000). Greenberg et al. (1999, 2002a) and Riley et al. (2000) argue that this impression is an artifact of the difficulty of recognizing older terrains that have been cut up by cracks, ridges and bands.

If the former interpretation is correct, Europa would appear to have changed in geological style over time, from
image resolution, however, led investigators to postulate that either the surface was very young, or that many of the mottled albedo features could be craters that had undergone viscous relaxation. For example, Lucchitta and Soderblom (1982) classified features observed in Voyager data into two categories: (1) craters several tens of km in diameter characterized by rims, central peaks and ejecta deposits, and (2) large, dark, flat circular spots \( \geq \)100 km in diameter (e.g., Tyre) on which lineaments appear to converge. Features in this latter category were interpreted to be relic, relaxed impact craters (Lucchitta and Soderblom 1982, Malin and Pieri 1986). Uncertainties about the identification and number of craters led to differences in the interpretation of surface ages (e.g., Lucchitta and Soderblom 1982).

**Galileo** targeting was designed to address many of the issues raised by the Voyager observations (Carr et al. 1995), including the nature of impact craters (e.g., Moore et al. 1998, 2001), implications for surface ages (e.g., Neukum 1997, Chapman et al. 1998, Zahnle et al. 1998), and insight into the outer layers of Europa (Turtle and Pizzo 2002, Schenk 2002). These topics are reviewed here and in Chapter 18.

### Morphology and Structure

Galileo data show that impact craters on Europa have the full range of features typical of craters on silicate bodies (Figure 15.12), such as bowl shapes in smaller examples (e.g., Govannan, \(~\)10 km in diameter), and flat floors, central peaks and massifs, terraces and extensive bright ray systems in larger examples (e.g., Manannán and Pywll, both in the 20–25 km diameter range). The larger features suspected to be of impact origin, such as Tyre, exhibit extensive secondary crater fields, excavated dark material, and an unusual interior structure.

Although there are similarities to craters on silicate bodies, fundamental differences are seen. For smaller craters, the transition from bowl-shaped to complex features occurs at about 5–6 km, with larger craters exhibiting central peak structures (Schenk 2002). Cilix (18 km diameter) is a good example of a complex crater; it has an elongate central peak complex, a flat floor, and terraced walls. The rim is polygonal and there is a reddish brown continuous ejecta deposit, suggesting that material different than that of the surface was excavated. A digital terrain model (Giese et al. 1999) shows that the floor is relatively flat and is at the same elevation as the surrounding terrain; the central peaks rise about 300 m above the floor, comparable to the elevation of the crater rim crest.

As crater diameter increases, some changes are observed. The 24 km diameter crater Pywll (Figure 15.12; Moore et al. 1998) has an extensive bright ray system extending \( \geq \)1000 km and a central circular dark spot that is a pedestal-like continuous ejecta deposit. The crater floor is relatively flat and extremely shallow (Geise et al. 1999). The proximal edge of the ejecta gives way to a bright braided pattern and radial rays of secondaries. A large peak complex \( \sim \)100–300 m high is offset from the center.

Secondary impact craters from Pywll range in diameters from \( \sim \)1 km to the limit of resolution (\( \sim \)10 m). High-resolution images in Conamara Chaos, some 1000 km from Pywll, show secondary craters from Pywll up to \( \sim \)500 m in
Figure 15.12. Impact craters >4 km in diameter on Europa, arranged in order of decreasing size. Italicized names are provisional (from Moore et al. 2001).
The onset diameter of central peaks and crater structure can be used to probe the nature of the outer layer(s) of Europa. Simple craters are interpreted to be too small to undergo significant collapse. Using the onset diameter for central peaks (>5 km) and estimates of transient crater depths, Moore et al. (2001) concluded that craters ~10–18 km in diameter (3–6 km deep transient craters) did not penetrate to a liquid layer and, thus, that the ice must be at least several km thick at the time of crater formation. Moore et al. (1998) and Schenk (2002) argue that the unusual morphology of impact structures ≥30 km in diameter, such as Callanish and Tyre, suggests that there was excavation to liquid water or possibly very warm ice at depth. Overall, the impact crater data appear to place the base of a solid ice layer at ≥20 km (Schenk 2002). The young age of the surface, as well as wide separation of Tyre and Callanish, suggest that this could represent a typical value for Europa for geologically recent times.

Galileo data show that the number of primary impact craters on Europa is small (e.g., Moore et al. 1998, 2001), suggesting a relatively young age, but there is debate about the absolute age of the surface. On the one hand, Chapman et al. (1998) suggested a mean age of ~10^7 to 10^8 years, while Neukum (1997) suggested that some surfaces on Europa are as old as 3.0–3.3 × 10^8, but could be as young as 10 million years. The differences are due to several factors, including assumptions about the age of Ganymede's oldest large basin (Gilgamesh), and uncertainties in the present-day comet impact rate in the jovian system. These issues are reviewed and discussed by Zahnle et al. (1998) and in Chapter 18, where it is argued, based on the dynamics of small body populations, that Europa's average surface age is nominally ~6 × 10^7 yr.

15.3.4 Global Tectonic Patterns and Stress Mechanisms

Diurnal Stressing

Europa’s orbit has a forced eccentricity of 0.01, maintained by the 3-body Laplace resonance with its neighboring satellites Io and Ganymede. The resultant tidal flexing of Europa occurs on the 3.55 day period of its orbital revolution, producing "diurnal" stresses. The magnitude of tidal flexing is a function of the satellite's interior structure. Total vertical deflection is expected to be only 1 m if the interior is solid, or 30 m if an icy shell is able to flex above a subsurface water layer, with only very weak dependence on the depth to the ocean (Moore and Schubert 2000). Consequently, stresses due to diurnal orbital flexing are expected to decrease markedly if there is no liquid layer at depth. Moore and Schubert (2000) show that an interior rigidity as low as 10^7 Pa could produce a significant tidal amplitude (18 m) without a subsurface ocean, but rigidity is so low that it would require the rock mantle to be substantially partially molten. The magnitude of the stresses for a floating ice shell is ~100 kPa = 1 bar (Helfenstein and Parmentier 1980, Cassen et al. 1982, Greenberg et al. 1998, Hoppe et al. 1999a).

Greenberg et al. (1998) modeled the diurnal stress surface pattern as it changes continuously, and recognized that the direction to Jupiter changes in Europa's reference frame as the satellite orbits. As a result, surface stresses at any given location change in magnitude and rotate during an orbital cycle, resulting in clockwise stress rotation in the southern hemisphere and counterclockwise in the northern hemisphere. This could account for some of the observed sur-
face features. For example, Galileo images show many cases of strike-slip offsets on Europa. Hoppa et al. (1999a) and Sarid et al. (2002) show a preferred direction of strike-slip offset in each hemisphere, with propensity for right-lateral strike-slip in the southern hemisphere, and left-lateral in the northern hemisphere. This could be explained by the rotation of diurnal stresses, with faults effectively “walking” in a right-lateral sense in the southern hemisphere, and in a left-lateral sense in the northern hemisphere (Hoppa et al. 1999a). Hoppa et al. (1999a) suggest that the observed strike-slip features formed at different longitudes with respect to Jupiter than currently occupied, and were shifted by nonsynchronous rotation (discussed below and in Section 15.4.2).

Diurnal stresses can explain the patterns of cycloidal ridges (Figure 15.13) and other cycloidal structures, such as some gray and dark band boundaries (Hoppa et al. 1999a, Greenberg et al. 2002a). If a fracture propagates across Europa’s surface at an appropriate speed (about 3 km h\(^{-1}\)), the stress orientation rotates during a fraction of the Europian day and the propagating fracture traces out the curvature of a single cycloidal arc. The diurnal stress then decreases below the critical value for fracture until the following orbit, when tensile stress again increases, and fracture propagation reinitiates, generating the next cycloidal arc. This model produces an excellent match to the shapes of cycloidal features if tensile failure occurs at a stress of \( \sim 25 \) kPa and propagation halts when stress is below \( \sim 15 \) kPa (all quite low values).

The Hoppa et al. (1999b) model explains several aspects of cycloidal structures shown in Figure 15.13. For example, the arcs of an individual cycloidal chain always show a consistent direction of convexity, while different chains can have opposite convexity directions. In the Hoppa et al. model, convexity direction simply depends on the fracture propagation direction relative to the sense of stress rotation. 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. As with the strike-slip “walking” model for strike-slip tectonism, longitudinal shifts must be invoked to account for present locations of cycloidal ridges relative to the local stress fields, suggesting nonsynchronous rotation of the surface since the formation of the visible cycloidal features (Hoppa et al. 2001).
Nonsynchronous Rotation

Europa's eccentric orbit causes its oscillating tidal bulge to lag behind the Jupiter-facing direction at perijove, creating a torque that acts to accelerate the satellite's rotation to slightly faster than synchronous (Goldreich 1966, Greenberg and Weidenschilling 1984, Ojakangas and Stevenson 1989b). We expect the rocky interior of Europa to maintain a permanent mass asymmetry sufficient to counter this torque, so it should be synchronously locked. However, Europa's icy near-surface layer could rotate nonsynchronously if it is decoupled from the rocky interior, as by a subsurface ocean.

Mechanisms and timescales of nonsynchronous rotation of Europa were considered by Greenberg and Weidenschilling (1984) and Ojakangas and Stevenson (1989b). With a permanent mass asymmetry, Greenberg and Weidenschilling (1984) envision a satellite locked in synchronous rotation with its "permanent" tidal bulge offset ahead of the Jupiter-facing direction at perijove, so that the average torque over an orbital period is zero. The permanent bulge could attempt to creep toward hydrostatic equilibrium (i.e., toward the Jupiter-facing direction), with the result that the satellite's surface will rotate nonsynchronously. Ojakangas and Stevenson (1989b) consider that nonsynchronous rotation could result from thermal adjustment of a floating ice shell. Predicted longitudinal variations in ice thickness (due to variations in tidal strain rate) will drive the ice shell slightly out of hydrostatic equilibrium. In attempting to move back toward hydrostatic equilibrium, the shell will rotate nonsynchronously (see Section 4.2).

The Greenberg and Weidenschilling (1984) model in its original form might operate if the subsurface is warm ice as opposed to an ocean, but viscous coupling with the presumably tidally locked rocky interior would certainly act to retard nonsynchronous rotation. In the Ojakangas and Stevenson (1989b) model, liquid water beneath the ice shell assures decoupling. However, nonsynchronous rotation is not a given in this model, because warm ice might flow rapidly at the base of the floating ice shell, eliminating the longitudinal shell thickness variations envisioned to drive the rotation (Stevenson 2000b). Alternatively, it is possible that nonsynchronous rotation is a permanent condition of the ice shell, and perhaps relatively rapid (Ojakangas and Stevenson 1989b). A lower limit of 10^4 years was 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).

Helfenstein and Parmentier (1985) and McEwen (1986b) predicted the stress pattern which should result from nonsynchronous rotation (Figure 15.14), based on a small eastward shift of Europa's surface relative to its fixed tidal axes (the pattern is independent of the amount). In the equatorial region, alternating zones of biaxial tensile and compressive stresses are predicted as the surface is stretched over, and falls off, the tidal bulge, respectively. Principal stresses are of opposite sign elsewhere across the satellite, with maximum differential stress occurring near the poles, and minima at the sub jovian and ant jovian points. Voyager global-scale lineaments were compared to this pattern by McEwen (1986b) and Leith and McKinnon (1996), who concluded that the best match of the nonsynchronous stress pattern occurred by considering a westward longitudinal shift in the locations of surface features relative to the fixed tidal axes (Figure 15.14). If the longitude of surface features is shifted westward (or equivalently, the tidal axes shifted eastward) to "back up" nonsynchronous rotation by ~25°, then lineament orientation is approximately perpendicular to the least compressive (greatest tensile) stress direction, as expected if the lineaments originated as tension or extension fractures. The implication is that Europa's major lineaments formed over a range of ~25°-50° of nonsynchronous rotation (because the stress pattern moves at half the nonsynchronous rotation rate), giving the observed best fit. Stresses generated by nonsynchronous rotation can be significant. Maximum stresses of ~0.14 MPa can be achieved per degree of rotation; thus, accumulated tensile stress can exceed the tensile strength of cold, solid ice upon ~12° of nonsynchronous rotation (Leith and McKinnon 1996).

Galileo color images of the northern high-latitude region of Europa's trailing hemisphere support and strengthen this argument. Imaging at near-infrared wavelengths discriminates older lineaments that were invisible to Voyager (Clark et al. 1998, Geissler et al. 1998a,b). Geissler et al. (1998b) categorized lineament age based on color characteristics, and found that lineament orientations have pro-

![Figure 15.14. Global-scale lineaments (solid) compared to the trajectories of least compressive stress (dashed) as predicted from nonsynchronous rotation (Helfenstein and Parameter 1985): (upper) the present-day position of surface features relative to the tidal axes, and (lower) surface features rotated "backward" (westward) in position by 25° of longitude. The shifted pattern offers a good overall fit in that the stress trajectories tend to be perpendicular to the observed lineaments. This suggests that Europa's prominent global-scale lineaments have formed over ~50° of nonsynchronous rotation. Light gray indicates zones in which both horizontal principal stresses are tensile, and dark gray indicates zones in which both principal stresses are compressive. Longitude is labeled in the tidal reference frame (after McEwen 1986b).](image-url)
gressively 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 formal stress mechanism, however, because the orientations of the most recent lineaments mapped by Geisler et al. (1998b, 1999) are better fit by diurnal stresses. It is plausible that diurnal stresses create some cracks, while nonsynchronous rotation opens those cracks into wider ridges and bands. The total amount of rotation inferred from the oldest global-scale lineaments is ~60° (Greenberg et al., 1998), slightly more than inferred from Voyager visible-wavelength imaging.

Galileo images show that Europa’s ridged plains are overprinted by ridges and ridge sets of various orientations. Cross-cutting relationships throughout the visible geologic record suggest evidence for at least one full rotation of Europa’s ice shell (Figueroa and Greeley 2000, Kattenhorn 2002). Considering Europa’s apparently young surface age of ~60 Myr (Chapter 18), the ice shell is usually assumed to rotate nonsynchronously at the same rate today as during the formation of its surface features (Hoppa et al. 1999c, 2001). However, it is also possible that Europa’s internal activity, including nonsynchronous rotation, has slowed, halted, or is episodic.

The equatorial region of isotropic tension west of the antijovian point, predicted by nonsynchronous rotation (Figure 15.14), correlates generally well with the zone of pull-apart bands originally recognized in Voyager images (Helfenstein and Parmentier 1980, Pieri 1981, Lucchita et al. 1981, Lucchitta and Soderblom 1982, Schenk and McKinnon 1989), and recognized in Galileo images to extend westward to ~250° longitude (Belton et al. 1996, Sullivan et al. 1998). More complex (but uncertain) stress sources are implied by the findings that this extensional zone is centered ~15° south of the equator, and that dark and wedge-shaped band opening directions have preferred orientations (Schenk and McKinnon 1989, Sullivan et al. 1998). Minor polar wander of Europa’s ice shell (Ojakangas and Stevenson 1989a) might influence patterns of band opening. Although this process is not indicated by global-scale lineament orientation as imaged by Voyager (Leith and McKinnon 1996), and would not be expected if the warm base of Europa’s ice shell can flow rapidly (Stevenson 2000b, cf. O’Brien et al. 2002), evidence for this process is suggested by the sense and distribution of strike-slip faults in Europa’s leading and trailing hemispheres (Sarid et al. 2002). A similar extensional region is predicted west of the subjovian point, but is not observed in Galileo images, perhaps because cracks formed in this region did not dilate (Hoppa et al. 2000). Zones of compressional stress are also predicted, centered 90° in longitude away from the extensional zones (Figure 15.14), and these could be regions where the ice shell fails in shear rather than in tension (Spaun et al. 2003, Stempel and Pappalardo 2002).

Greenberg et al. (1998, 2002a) recognized that the rapidly changing pattern of diurnal stresses could add to the longer-term pattern of nonsynchronous stress to affect the overall surface fracture pattern. This additive stress pattern changes throughout the European day, with tensile stresses peaking at different locations at different positions along the orbit. Tensile failure is expected to occur when the additive

| Table 15.1. Properties of Europa.\(^a\) |
|-----------------------------|-----------------------------|
| Radius, \( R \)             | 1560.7 (± 0.65) km          |
| Mass, \( M \)               | 4.79982 (± 0.00062) \( \times 10^{22} \) kg |
| \( GM \)                    | 3292.72 (± 0.05) \( \text{km}^3 \text{s}^{-2} \) |
| Surface gravity             | 1.315 (± 0.001) m s\(^{-2} \) |
| Mean density \( \rho \)     | 3014 (± 4) kg m\(^{-3} \)   |
| \( C_{22} \)                | 132.2 (± 2.1) \( \times 10^{-6} \)       |
| \( C/\text{MR}^2 \)         | 0.348 ± 0.003               |
| Distance from Jupiter       | 6.709 \( \times 10^{8} \) km |
| Orbit period                | 3.551 day                    |
| Forced eccentricity         | 0.0093 (variable)           |
| Inclination                 | 0.464°                       |


The effects of nonsynchronous stress and diurnally varying stress overcome the lithosphere’s tensile strength. In some regions analyzed, the net effect on the expected fracture pattern is a 10° westward shift of the nonsynchronous pattern relative to the tidal axes, apparently explaining why some surface features would better fit the nonsynchronous stress pattern if the lineaments are shifted 10° eastward of their present locations (Greenberg et al. 1998).

15.4 THE INTERIOR

The interior of Europa exerts a primary control on the satellite’s geological evolution and astrobiological potential. Results from the Galileo mission and theoretical studies have greatly enhanced our understanding of the interior with respect to that of the immediate post-Voyager era (Malin and Pieri 1986).

15.4.1 Gravity and Magnetic Data

Europa’s density and icy surface suggest that it is primarily a silicate body that is partly differentiated. Early models ranged from an anhydrous rock and metal core covered with ~100-km-thick layer of ice (Consolmagno and Lewis 1976, Fanale et al. 1977, Cassen et al. 1982), to a partly hydrated silicate interior (serpentine rich) with only a thin carapace of ice (Ransford et al. 1981). Galileo encounters enabled refined determinations of Europa’s mass, radius, and density (Table 15.1), and measurements of its second-degree gravitational harmonic coefficient \( C_{22} \) to determine a normalized moment of inertia of 0.346 ± 0.005 (Anderson et al. 1998) (the slightly different values in Table 15.1 are normalized to the radius in Davies et al. 1998). This result is based on a combined analysis of passes E4, E9, E11, and E12, and assumes that the second-degree gravity field is due entirely to hydrostatic distortion of Europa’s figure by rotation and jovian tides. Because independent confirmations of close-to-hydrostatic conditions exist for Io (Anderson et al. 2002) and Ganymede (Anderson et al. 1996) from equatorial and polar passes, the same is assumed for Europa.

Europa’s relatively low moment of inertia implies a differentiated interior; for comparison, the values for the Earth and Mars are 0.334 and 0.366, respectively (Fölkner et al. 1997). More importantly, the thin-ice layer model of Ransford et al. (1981) can be rejected. Although an interior model
with a thin (<25 km thick) ice shell can be proposed in order
to fit both the density and moment-of-inertia constraints,
the non-ice interior would have to consist of a metallic
core and an underdense (<3000 kg m⁻³) mantle. Cosmo-
chemically, such underdense compositions are only possible
through hydration, but this is not consistent with the high
temperatures, melting, and differentiation that are involved
in the formation of a metallic core (Anderson et al. 1998).

Constraints on the internal structure can be determined
from gravity data through models. Figure 15.15 shows the
ice layer, mantle, and core size in a three-layer model as
a function of two representative core densities (Fe and Fe–
FeS composition). The ice shell density is assumed to be
1050 kg m⁻³, which is plausible for an ice or water shell
that contains some salts or sulfates (e.g., the density of cold
polar seawater is 1028 kg m⁻³). From such models, Anderson
et al. (1998) conclude that the ice shell is 80 to 170 km
thick. Although they do not favor two layer models that lack
metallic cores, because these require a rock interior density
(≈3800 kg m⁻³) greater than that of bulk Io (3530 kg m⁻³),
such interior densities are cosmochimically possible (Consol-
in shell density, from pure ice to sulfate rich (e.g., Kargel
et al. 2000), also broaden the range of ice shell thicknesses.
Of the models in Figure 15.15, the Fe–FeS model is likely
closer to reality, based on thermal evolution considerations.
Consequently, the metallic core is probably about half the rock
interior in radius (Anderson et al. 1998). For mantle densi-
ties similar to Earth’s upper mantle (∼3400 kg m⁻³), the
ice shell is probably >100 km thick (Chapter 13). Note that
the gravity models are not sensitive to the physical state of
the “ice” layer, i.e., whether it is solid or liquid beneath the
surface.

Galileo magnetometer data provide additional con-
straints on models of the interior (Khurana et al. 1998,
Kivelson et al. 1999, 2000, Zimmer et al. 2000). These data
show evidence of an induced dipole magnetic field con-
sistent with the presence of a salty ocean (Chapter 21).
Jupiter’s magnetic field is inclined to its rotation axis, and
the Galilean satellites (in Jupiter’s equatorial plane) expe-
rience a time-varying magnetic field with a synodic period
of 11.23 hr at Europa. Europa responds to this field as a
nearly perfectly conducting sphere, generating electrical cur-
rents that create a magnetic field that opposes and cancels
the time-varying component in its interior, but exterior to
Europa this induced field is manifested as a time-varying
dipole.

On the timescale of a given spacecraft pass, this time
variation is not seen, but multiple flybys show the proper
dipole signature, oriented principally in the equatorial plane
and in the appropriately different directions for each en-
counter (Kivelson et al. 2000). There is little doubt that
this observed behavior indicates a conducting layer within
Europa (cf. Stevenson 2000a). Because the ionosphere of Eu-
ropa is insufficiently conductive to carry the required cur-
rrents, the conductive layer must be within the body of Eu-
ropa (Zimmer et al. 2000).

Figure 15.16 illustrates the characteristics of a conduct-
ing shell model. The normalized amplitude A and phase lag
of the induced response are plotted as a function of shell
thickness and conductivity. A perfect conductor will have
an A = 1 and zero phase lag. A real conductor will have

\[ A < 1 \text{ and a finite phase lag. The magnetometer data indi-
}

\[ c_{22} \text{ as the conductivity source because it would be too deep.} \]

On the other hand, the conducting layer could lie at
the top of the rock mantle if the ice layer is thin, but the
rock conductivity would have to exceed ∼0.2 S m⁻¹ (Zim-
mer et al. 2000). These conductivities are unrealistically
high for ordinary rock and would require a hot, saline pore
fluid at volume fractions in excess of ∼10% (or something
exotic, such as an interconnected network of graphite or
pyrrhotite; Grant and West 1965, p. 398, Stevenson 2000a).
This amount of hot, reactive fluid is not stable at the pres-
sures characteristic of the putative conducting layer (>0.1
GPa), and should be largely expelled to the ice layer. In con-
trast, very hot (>800°C), electrolyte-bearing (for normal
porosities) or even partially molten rock could have the re-
quired conductivity (Hermance 1995, p. 201), but such high
temperatures in the outer 100 km of the rock mantle would
imply high heat flows and, as Zimmer et al. (2000) argue,
would probably melt the ice layer above.

The source of high conductivity is thus most plausibly
within the ice layer. The inferred 80 to 170 km thickness
for the ice layer requires that its average conductivity ex-

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tidal potential and associated deformation of electrolyte (e.g., sulfates, alkali salts); this ocean does not thicken beneath its ice shell and that the ocean contains an with both Io and Ganymede (the Laplace resonance), which suggest that Europa has an ocean at least several kilometers distributed at porosities > 10%, as discussed by Zimmerman et al. (2000). This eccentricity creates variations in the gravity and magnetometer data. As a complement to tidal strains on Europa, which are small. Moreover, with this mechanism, there does not appear to be any observed interval of transient creep (Durham et al. 2001), so it should be applicable to modeling the response to tidal strains on Europa, which are small.

Using the rheological parameters of Goldsby and Kohlstedt (2001) the Maxwell time of ice on Europa is

$$\tau_M = 77 \text{ hr} \left( \frac{d}{5 \text{ mm}} \right)^{1.4} \left( \frac{\sigma}{0.1 \text{ MPa}} \right)^{-0.8} \exp \left( \frac{270}{T} - 1 \right)$$

For an average stress level $\sigma$ of 0.1 MPa (see below) and a grain size $d$ of 5 mm (comparable to polar glacier ice), the Maxwell time at the melting temperature ($T \approx 270 \text{ K}$) is close to the tidal forcing period. This estimate is sensitive to grain size and stress level, but the most important variable is the rheology. Near the melting temperature ($>258 \text{ K}$), the viscosity of ice (and thus the Maxwell time) decreases with respect to GBS or GSS creep by an order of magnitude or more due to premelting on grain boundaries (De La Chappelle et al. 1999, Goldsby and Kohlstedt 2001). Consequently, Europa’s ice layer is probably mostly heated near its base (Ojakangas and Stevenson 1989b), which maximizes the influence of tidal heating on the layer as a whole; indeed in conducting ice tidal heating models, the resulting temperature gradient is exponential in the subsurface with tidal heating, peaking just above the ice-water interface (Chyba et al. 1998). Moreover, it implies that warm ice elsewhere could be similarly tidally heated (McKinnon 1999, Sotin et al. 2002). In contrast, the Maxwell times in Europa’s much more viscous rock mantle cannot approach the tidal period unless the rock is partially molten.

In summary, Galileo gravity and magnetometer data suggest that Europa has an ocean at least several kilometers thick beneath its ice shell and that the ocean contains an electrolyte (e.g., sulfates, alkali salts); this ocean does not need to be as conductive (or salty) as Earth’s ocean.

15.4.2 Tides and Tectonics

Europa’s complex geology reflects substantial processes originating in the interior. Europa is in a mean-motion resonance with both Io and Ganymede (the Laplace resonance), which currently forces Europa’s eccentricity $e$ of 0.01 (Greenberg 1982, Peale 1980). This eccentricity creates variations in the tidal potential and associated deformation of ∼3% on an orbital period of 85.2 hr, and the dissipation of the resulting strain energy heats the interior.

Tidal dissipation in planetary bodies can occur by many mechanisms and on several scales, e.g., from viscoelastic heating on the grain scale, to solid friction along faults, to turbulence at liquid–solid boundaries. Viscoelastic heating depends on the rheology of the material and peaks when the period of forcing equals the Maxwell time $\tau_M = \eta \mu = \sigma / 3\mu$, where $\eta$ is viscosity, $\mu$ is the shear modulus (3.52 GPa for solid ice; Gammon et al. 1983), $\sigma$ is stress, and $\varepsilon$ is strain rate. For water ice, the dominant low-stress deformation mechanism should be dislocation-accommodated grain-boundary sliding (GBS, or superplastic creep), unless the grain sizes are large (Goldsby and Kohlstedt 1997, 2001). Durham et al. (2001) and Durham and Stern (2001) refer to this more generically as grain-size-sensitive, or GSS, creep. Moreover, with this mechanism, there does not appear to be any observed interval of transient creep (Durham et al. 2001), so it should be applicable to modeling the response to tidal strains on Europa, which are small.

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Diurnal tides also affect the outer, brittle part of the ice shell and are responsible for tectonic processes. Europa’s equilibrium tide and rotation combine to yield a hydrostatic triaxial figure with $a$, $b$, and $c$ axes whose lengths differ from Europa’s average radius by 1.87, −0.53, and −1.33 km, respectively, where $a$ is the tidal axis (oriented toward Jupiter), $c$ is the rotational axis (perpendicular to the orbital plane), and $b$ is the intermediate axis (orthogonal to the other two).

As Europa’s distance from Jupiter varies during its orbit, the tidal potential rises and falls by 3e, whereas the rotation of Europa’s mass is close to invariant (the forced libration is ±8 × 10−4; Peale 1977, Eq. 22). Thus, the tidal distortion alone fluctuates, reaching its maximum and minimum at perijove and apojove, respectively, and the $a$ axis of
a hydrostatic (fluid) Europa rises and falls, respectively, by 48 m. The rigidity of the body of Europa offers resistance to this relatively high-frequency deformation, however, and Yoder and Sjogren (1996) and Moore and Schubert (2000) calculate that the \( a \)-axis change is reduced to \( \pm 30 \text{ m} \). A floating, thin ice shell (up to a few tens of km) will conform to this shape change (e.g., Ojakangas and Stevenson 1998b). The resultant elastic stresses in the shell are given by the solution to the biaxial distortion of a thin shell (Melosh 1977)

\[
\sigma_{\theta\theta} = \frac{1}{3} (f - f_0) \mu \left( \frac{1 + \nu}{5 + \nu} \right) (5 + 3 \cos 2\theta) \\
\sigma_{\phi\phi} = -\frac{1}{3} (f - f_0) \mu \left( \frac{1 + \nu}{5 + \nu} \right) (1 - 9 \cos 2\theta)
\]

(15.2) where \( \sigma_{\theta\theta} \) and \( \sigma_{\phi\phi} \) are the meridional and azimuthal stresses referenced to the \( a \)-axis. The parameter \( \theta \) is the colatitude, \( f - f_0 \) is the difference between initial and final flattenings, \( \mu \) and \( \nu \) are the shear modulus and Poisson's ratio, respectively, of the shell (\( \nu = 0.325 \) for solid ice; Gammon et al. 1983), and compression is defined as positive. The change in flattening can be written \( 3\epsilon_\gamma \), where \( \gamma \) is a factor that accounts for the flattening reduction due to rigidity (\( \approx 0.62 \)), and \( f_0 \) is the tidal flattening (negative for elongation; see Leith and McKinnon 1996, p. 394). Evaluation of these parameters indicates that at perijove and apojove maximum stresses do not exceed \( \approx 0.08 \text{ MPa} \).

Between perijove and apojove, the tidal potential passes through its average value, but because Europa's instantaneous orbital angular velocity and spin rate are not exactly equal, the tidal potential and \( a \) axis are offset by 2e radians (Greenberg et al. 1998). The resulting elastic stresses in the shell are determined by the partial relaxation of the tidal bulge and its reimposition with a 2e offset, equivalent to nonsynchronous rotation of the shell by the same amount (Greenberg et al. 1998). The stress generated in this case (Leith and McKinnon 1996) is

\[
\sigma_{\theta\theta,\phi\phi} = \pm \gamma f_0 \left( \frac{1 + \nu}{5 + \nu} \right) \sin \Omega
\]

(15.4) where \( \Omega \) is the nonsynchronous rotation angle (2e), or 0.10 MPa.

Thus, diurnal eccentricity stresses are small (\( \leq 1 \text{ bar} \)), but they have been invoked to explain ridge-building (Greenberg et al. 1998, 2002a, Gaidos and Nimmo 2000, Nimmo and Gaidos 2002), incremental strike-slip motion (Hoppa et al. 1999a), wedge-shaped band dilation (Tufts et al. 2000), cycloid ridge propagation (Hoppa et al. 1999b), and even eruption of ocean water (Greenberg et al. 1998, 2002a). With regard to the latter, 0.1 MPa of tension can open a single crevasse no deeper than \( \pi \sigma/2pg \approx 150 \text{ m} \), where \( \rho \) is the ice density and \( g \) is surface gravity (e.g., Weertman 1971). Fluid-filled crevasses can propagate upward from the base of the shell to much greater heights if sharp cracks can be initiated in the soft basal ice (Crawford and Stevenson 1988), as can be seen by replacing \( \rho \) in formula above with \( \Delta \rho \), the density difference between the ice and ocean water. For the minimum \( \Delta \rho \) given by Kargel et al. (2000) of \( \approx 64 \text{ kg m}^{-3} \), for a eutectic sulfate ocean and shell, basal crevasses may reach heights of \( \approx 2 \text{ km} \) driven by diurnal stresses alone.

The diurnal tides and stresses are much reduced if Europa's shell is mostly or completely frozen. The tidal variation in the \( a \)-axis in this case would be \( \leq 1 \text{ m} \) (Yoder and Sjogren 1996, Moore and Schubert 2000), which implies a >30-fold reduction in stresses and a <(30)2 \( \approx 10^8 \) reduction in tidal heating. Most important, if the kinematic requirements for diurnally varying stress in generating cycloid cracks (Hoppa et al. 1999b) are accepted, in which each cycloid arc is generated within one Europa day, then it is inconceivable that these could be met by <3 kPa tidal stresses driving the propagation of <5-m-deep surface cracks across hundreds of km of uneven terrain. The suggestion that cycloid ridges evolve from cycloid cracks, and the stress cycle needed to generate them, are powerful arguments for the presence of an ocean decoupling the shell from the interior.

Diurnal tidal stresses in a floating ice shell can propagate tension cracks, but if such cracks are to reach the surface from the base of the shell or penetrate from the surface to ductile ice below, one or more of the following are required: (1) a more sophisticated crack mechanics (e.g., the gas exsolution model of Crawford and Stevenson (1988); (2) a relatively thin (<3-4 km thick) ice shell, now or in the past; and (3) additional sources of stress to fracture the shell.

One additional stress source is nonsynchronous rotation. With internal dissipation, the diurnal tidal response is retarded in time or delayed in phase. The phase lag, expressed as an angle, is usually small and depends on the magnitude of the dissipation (i.e., the tidal \( Q \)). The average torque on Europa is weighted towards perijove, however, where the tidal response trails the tidal potential and is positive (i.e., acts to spin up the satellite). Normally, for solid satellites in 1:1 spin–orbit resonance, this average torque is balanced by an opposite torque on the permanent tidal bulge, which is slightly offset in the other direction at perijove, ahead of the tidal potential (e.g., Yoder 1979). The position of the permanent bulge with respect to the body of Europa in the long term is determined by the non-hydrostatic mass distribution within the satellite, such that the \( a \)-axis and \( c \)-axis are also the minimum and maximum moment-of-inertia axes, respectively, of the undistorted satellite. As long as the non-hydrostatic contribution of the rock interior to \( C_{22} \) is \( >10^{-6} \), the interior should stay tidally locked in the 1:1 spin–orbit resonance, and for so-called Darwin tides, the minimum non-hydrostatic \( C_{22} \) for tidal locking is 10–100 times less (Greenberg and Weidenschilling 1984, Schenk and McKinnon 1989, Ojakangas and Stevenson 1998b). The non-hydrostatic stresses necessary to support these interior mass anomalies are \( \approx 10\% \) of those supported in the lunar lithosphere (Solomon 1986), or \( \approx 10 \text{ MPa} \). Unfortunately, there is no independent measure of the non-hydrostatic mass distribution within Europa, but if it is insufficient, then the permanent bulge will seek to relax viscously to the position of the potential minimum at perijove (Greenberg and Weidenschilling 1984). Because it can never catch up, the permanent bulge slowly creeps westward through the body of Europa, and Europa effectively rotates slightly faster than synchronously (Greenberg and Weidenschilling 1984).

Even if the rock interior is tidally locked, a floating ice shell is subject to the same subtle interplay of torques (e.g., for shell thicknesses <30 km, the tidal lag angle is <0.5°; Moore and Schubert 2000), and the shell can undergo independent nonsynchronous rotation. It is unlikely that a non-hydrostatic distribution of ice phases could permanently stabilize a floating ice shell with respect to the tidal axis given viscous creep at its base (cf. Stevenson 2000b). In the tidally heated shell model of Ojakangas and Stevenson...
(1989b), however, variations in tidal heating with latitude and longitude imply maximum shell thicknesses and, thus, the minimum moment of inertia of the shell along the a axis, appropriate for stability. Even in this case, the maximum thickness will be offset dynamically from the potential that ultimately creates it, leading to nonsynchronous rotation on a timescale governed by the continuous thermal reequilibration of the shell.

For their estimated shell thickness \( b \) of \( \sim 15\text{–}25 \text{ km} \), Ojakangas and Stevenson (1989b) predicted a nonsynchronous rotation time of \( \sim 10 \text{ Myr} \), consistent with the lower limit of Hoppa et al. (1999c) but inconsistent with the upper limit of Hoppa et al. (2001). This timescale goes as \( \sqrt{b} \) thus, a 10-km-thick shell could rotate in 2.5 Myr. The shell could rotate at an even faster rate (governed by viscous processes), and the tidal heating would be zonally averaged, leading to a maximum and uniform shell thickness on the equator. Regardless of the nonsynchronous rotation rate, the possible stresses in Europa’s lithosphere are large. For a full 90° of rotation and \( \gamma = 0.62 \), the maximum tensile stresses from Eq. (15.3) reach 8.1 MPa, adequate in principle to open surface tension cracks to a depth of \( > 10 \text{ km} \) (Leith and McKinnon 1996). More likely, small amounts of nonsynchronous rotation stress could enhance the rise of fluid-filled cracks from the base of the shell (Greenberg et al. 1998), and such stresses would not necessarily overwhelm the diurnal stress pattern variations suggested for some of the tectonics.

Ojakangas and Stevenson (1989a,b) noted that in their nominal (trialvaloid) tidal heating model the polar regions of the shell were sufficiently thick due to low surface temperature, that the polar axis was actually the intermediate moment-of-inertia axis, rather than the minimum as required by stability. Thus, a dynamic driver could exist for true polar wander, attempting an interchange of the \( b \) and \( c \) axes by means of rotation about the \( a \) axis. Whether this polar wander actually occurs depends on the dissipative properties of the shell; it may be episodic, occur continuously as the shell thickness adjusts to the evolving position-dependent tidal heating, or not occur at all. The shell stresses it creates are identical in form to nonsynchronous stresses, except that the \( a \) axis rather than the \( c \) axis is invariant, and the stress levels are reduced by a factor of 3 for a given rotation angle (Leith and McKinnon 1996).

In summary, the tides and torques raised on the ice shell lead to several stress sources for tectonic activity. These operate on various timescales and stress levels, which could explain the variety of tectonic surface features. In detail, however, much remains to be understood, especially in regard to how much nonsynchronous rotation has occurred through time and whether polar wander has occurred.

15.4.3 Tidal Evolution

The Laplace resonance supplies the orbital eccentricity that drives Europa’s tectonics and heats its interior. A leading model for its initiation (Yoder, 1979, Yoder and Peale 1981) postulates that differential tidal expansion first caused Io to move into the 2:1 mean-motion resonance with Europa, after which the pair was tidally driven outward until Ganymede was captured into the 2:1 resonance with Europa as well. In this model, Europa’s eccentricity was much smaller (0.0014) when it was solely in the 2:1 resonance with Io, and it is only when the Laplace resonance formed that modern values of Europa’s eccentricity \( e \sim 0.01 \) were achieved.

The Laplace resonance can be illustrated through the pair-wise relationship between satellite mean motions, \( n_i \), and the drift rate between conjunctions,

\[
2n_i - n_j = \omega_i \tag{15.5}
\]

\[
2n_j - n_i = \omega_j \tag{15.6}
\]

where \( n_i, n_j, \) and \( \omega_i \) refer to the mean motions of Io, Europa, and Ganymede, respectively. Currently, \( \omega_i \) and \( \omega_j \) are small and equal (time averaged), and equal to the orbital precession rate of either Io or Europa. Malhotra (1991) and Showman and Malhotra (1997) show that before Io and Europa entered the 2:1 mean-motion commensurability, all three satellites may have been temporarily captured into low order Laplace-like resonances characterized by \( \omega_i/\omega_j = 1/2, 3/2, \) or 2, after which the satellites evolved into the present \( \omega_i/\omega_j = 1 \) (Figure 15.17). Eccentricities are enhanced in these temporary resonances for both Europa and Ganymede, with Europa attaining \( e \) values that varied but which could have averaged up to \( \sim 0.01 \) (somewhat greater as the \( \omega_i/\omega_j = 3/2 \) and 2 resonances were exited). Thus, Europa’s modern eccentricity and level of tidal heating could date from this era in solar system history.

Alternatively, Greenberg (1982, 1987) argued that the Laplace resonance is primordial, and that the satellites were originally “deeper” in the resonance, meaning that their primordial forced eccentricities and heating rates were higher. Peale (1999) criticized this scenario as ad hoc, in that it was unlikely that the satellites simply formed in the resonance. In the model of Canup and Ward (2002), however, satellite-disk torques can cause substantial inward satellite migration. Because more massive Ganymede would drift faster, they speculate and Peale and Lee (2002) demonstrate that the Laplace resonance might have been assembled from the outside-in during this earliest epoch. Furthermore, Peale and Lee (2002) show that after satellite migration ceases, tidal dissipation causes the system to evolve to the current orbital configuration.

Unfortunately, Europa’s early history is not preserved in the geological record. Moreover, tidal heating rates could vary through time, even during the modern epoch. Fundamentally, there is a strong nonlinear feedback between tidal heating and orbital eccentricity: high dissipation (high \( T \) and low \( Q \)) lowers \( e \), which lowers dissipation, causing \( T \) to decrease and \( Q \) to increase, which allows \( e \) to be tidally pumped back up, causing \( T \) to increase and \( Q \) to decrease, and the cycle repeats. This argument, first articulated by Greenberg (1982) for Io and modeled in detail by Ojakangas and Stevenson (1986), suggests that Io’s eccentricity could vary up to \( \sim 0.01 \) from its present 0.0041 on a \( \sim 100 \text{ Myr} \) timescale. In the resonance, the variation in Io’s \( e \) is accompanied by a variation in \( n_1 \). The variation in \( n_1 \) must also drive variations in \( n_2 \) and \( n_3 \) to maintain the overall Laplace resonance lock, which can be expressed as \( n_1 - 3n_2 + 2n_3 \equiv 0 \) (subtracting Equation 15.6 from 15.5). If the Ojakangas and Stevenson (1986) model is correct, the implications for Europa could be significant. From Yoder and Peale (1981), a doubling in Io’s eccentricity could similarly result in a doubling of Europa’s \( e \) to \( \sim 0.02 \), which would increase the present tidal heating rate by a factor of 4. We note that
the preserved geological history of Europa overlaps the last eccentricity tidal heating maximum for Io, as modeled by Ojakangas and Stevenson (1986). This could be of relevance to arguments for secular changes in Europa's geological expression.

15.4.4 Thermal Evolution

Europa presumably formed within a flattened, rotating gaseous subnebula about Jupiter as Jupiter formed (e.g., Pollack and Fanale 1982, Stevenson et al. 1986, Peale 1999). The details of formation are under study (Canup and Ward 2002, Mosqueira and Estrada 2003a,b, Chapter 13), but because Europa is rock+metal-rich, its differentiation in terms of ice from rock+metal was assured. The melting and separation of water in a hypothetical primordial Europa of mixed ice and rock+metal is a given even if only long-term radiogenic heating is assumed. Based on gravity data, the ice/water volume within Europa is 15–30%, much less than the critical 40% or more needed for convective regulation to maintain internal temperatures below the ice melting point (Friedson and Stevenson 1983).

Much of the rock might have been initially hydrated, either in the jovian subnebula (Prinn and Fegley 1981) or by aqueous alteration within Europa, but hydration has not survived in bulk, given the gravitational evidence for metallic core formation or (in the absence of a metallic core) a dense, metal-bearing interior composition. It is not completely clear if radiogenic heating was sufficient to drive the water out of the hydrated minerals (Ransford et al. 1981). Europa, however, could have been more strongly heated during accretion (Lunine and Stevenson 1982, Stevenson et al. 1986) or by tidal activity.

Once the rock+metal interior separates, its internal temperature is no longer buffered by the presence of the ice, and temperatures can rise by radiogenic heating alone to approach the silicate solidus. Approaching the solidus from below means that iron can melt and drain downward before silicate melting occurs, provided that the presence of sulfur lowers the melting point of the resulting metallic alloy (McKinnon 1996). Otherwise, solid-state convection of the dominant rock fraction will regulate the internal temperatures to remain below the relatively high melting point of pure iron, and a metallic core will never form (e.g., Stevenson et al. 1983). The only alternative is for Europa's internal temperatures to have been so overdriven by tidal heating that sufficient rock melting occurs, such that metal-rich solids segregate to the center of the satellite. This is unlikely for present rates of tidal heating, but the effects of excursions to high eccentricity and tidal heating discussed earlier have yet to be evaluated.

Figure 15.17. (a) Example of tidal evolutionary paths for Io, Europa, and Ganymede displayed in a $\omega_1$-$\omega_2$ diagram. The initial slopes are flatter because Io's outward evolution is initially the greatest of the three; (b) time evolution of Europa's eccentricity for the example of temporary capture into the $\omega_1/\omega_2 = 3/2$ resonance (arrowed path in (a)); $Q_J$ is Jupiter's dissipation factor, (Peale 1999). In this numerical calculation Io and Europa were relatively dissipative, while Ganymede was not, and the system "jumped" into the Laplace resonance ($\omega_1 = \omega_2$) when Io's dissipation was lowered by a factor of 3. The state of the system at the end of the integration is close to that presently observed. After Showman and Malhotra (1997).

Figure 15.18 illustrates models by McKinnon (1996), updated for internal structures that satisfy both density and moment-of-inertia constraints. The models assume early formation of a rock+metal interior and radiogenic heating based on the $U/Si$, $Th/Si$, and $K/Si$ abundance ratios in carbonaceous chondrites (Mueller and McKinnon 1988). The interior heats during the first Gyr and reaches temperatures comparable to the Fe–FeS eutectic melting temperature, but still below temperatures characteristic of solid-state convection in rock. Once sufficient heat is released to supply the latent heat of melting of an Fe–FeS core, at ~1.5 Gyr, the core is considered formed, and the gravitational potential energy released compensates for the latent heat consumed. As the temperature continues to rise, solid-state convection in the rock mantle sets in (based on an olivine rheology), indicated by the flattened temperature profiles in the lower mantle. The flattened profiles in the core are simply due to the high thermal conductivity of molten Fe–FeS. Later, radiogenic heating decreases and the mantle cools and becomes conductive (Figure 15.18a). As core temperatures approach the Fe–FeS eutectic curve, refreezing might occur, but the details depend on the precise core composition and are not addressed.

The potential effect of tidal heating on thermal evolution is shown in Figure 15.18b, in which tidal heating is proportional to the ratio $k/Q$, where $k$ is the second-degree potential Love number. A substantial liquid core and the
oceanic layer raise and lower $k$, respectively, relative to that of a solid, uniform elastic sphere (cf. Cassen et al. 1979, Ross and Schubert, 1987), but the combined effect is an increase in $k$ for the rock-metal interior. In contrast, estimating the tidal dissipation factor $Q$ is difficult. It is generally supposed that $Q \approx 100$ is a good measure of solid body dissipation, although the lunar $Q$ at the monthly libration period is $\approx 25$, a fact exploited by Squyres et al. (1983) in their early study of Europa’s possible shell thickness (cf. Williams et al. 2001). $Q = 25$ is used in Figure 15.18b, where the tidal heating begins with core formation. The long-term decline in radiogenic heating is more than compensated by this level of tidal heating. The metallic core remains at temperatures well above the Fe-FeS eutectic temperature is from Fei et al. (1997). Fe-FeS eutectic melting temperature is from Fei et al. (1997).

Tidal dissipation in this calculation is evenly distributed in the mantle, and the heating rate is constant. Although these assumptions are simplistic, the calculations demonstrate that for plausible choices of physical parameters, Europa could have a thermally and volcanically active rock mantle at the present time. A corollary is that simple, Newtonian rheologies, it was assumed that maximum heating was at the very bottom of the shell. Based on the discussion of Maxwell times, it is plausible that the maximum heating occurs within the shell, which makes shell thicknesses $\leq 15$ km more likely for present-day conditions. On the other hand, frictional dissipation along faults or locally in the shell (e.g., Nimmo and Gaidos 2002) will tend to concentrate heat, which might lead to a warmer, thinner crust locally, but also reduce the strain energy available elsewhere, allowing the shell elsewhere to thicken (Stevenson 1996).

McKinnon (1999) considered the initiation of convection in the ice shell, based on scaling of convection in temperature-dependent fluids, low-stress ice rheologies, and the current tidal strain field, and found that convection could start in shells $< 20$ km thick if the ice grain size were $\sim 1$ mm or less. Convective overturn in shells as thin as 5–10 km is conceivable as well, but the grain size and basal viscosity would have to be very low ($\sim 100 \mu m$ and a few $10^{12}$ Pa s). Even if convection initiates, there might be no steady state because convective ice should be relatively hot and dissipative at tidal frequencies (McKinnon 1999). Hussman et al. (2002) propose, based on a different rheological model, equilibrium solutions for thick ($\sim 35–40$ km) shells, with convecting sublayers if the melting-point viscosity is sufficiently low. As discussed in Section 15.3.1, the sizes, spacings, and morphologies of pits, spots, and domes suggests some form of convective upwelling.
Unfortunately, the thickness and thermal state for Europa’s shell cannot be specified from theoretical arguments alone, because there is a lack of information on parameters such as grain size and the roles of viscous and frictional dissipation, and because the history of the shell and interior strongly determines the shell’s present behavior. Nonetheless, models suggest shell thicknesses that vary with location and time, and which are consistent with the tectonic, volcanic, and other morphologies observed. Uncertainties in our knowledge of shell and ocean properties have a direct bearing on the prospects for life on Europa, discussed next.

15.5 ASTROBIOLOGY

Life as we know it depends on liquid water, a suite of “biogenic” elements (e.g., carbon, but others discussed below), and a source of energy (on Earth, either sunlight or chemical disequilibrium, which may or may not be traceable to sunlight). The suggestion for a subsurface liquid water ocean on Europa, coupled with discoveries of an extensive subsurface biosphere on Earth (Gold 1992, Whitman et al. 1998) combine to make Europa’s putative subsurface ocean one of the most promising sites for exobiology in the solar system.

We note that if the origin of life requires direct access to the enormous free energy available from the Sun, then life on Europa would not exist unless it arose in a possibly brief, early intense greenhouse stage, and quickly adapted to life beneath an ice cover, or if it reached Europa through the successful interplanetary transfer of microorganisms from elsewhere. Although such a transfer is feasible between Earth and Mars (Mileikowsky et al. 2000), it is far more difficult with Europa. Objects will strike Europa at velocities unimpeded by any significant atmosphere, making high temperature shock heating much harder to avoid (Pierazzo and Chyba 2002).

Certain prebiotic chemical processes under hydrothermal conditions might have been important in the origin of terrestrial life (Wächtershäuser 1988, Cody et al. 2000), but it is possible that other required compounds included molecules such as sugars (Weber 2000) which could have required a surface origin. The terrestrial origin of life remains too poorly understood (Chyba and McDonald 1995) for firm conclusions to be drawn. The investigation of Europa could provide insight into the importance of different environments to the origin of life.

15.5.1 Biogenic Elements

Besides liquid water, life requires a suite of biogenic elements, including carbon, hydrogen, oxygen, nitrogen, phosphorus, sulfur, and others. Spectroscopic observations provide some data on the composition of the upper surface of Europa, as reviewed in Section 15.2. Various organic groups, such as C=O and C–H, have been detected on Callisto and Ganymede, and there are suggestions of these compounds on Europa.

However, estimates of abundances of biogenic elements rely on Europa formation models, some of which suggest that Europa should be volatile-rich. But even if Europa formed extremely volatile-poor, comet impacts should have delivered substantial quantities of biogenic elements (e.g., ~10^{12} kg C) over the age of the solar system (Pierazzo and Chyba 2002). Thus, the availability of biogenic elements seems unlikely to be an impediment to the existence of life on Europa.

15.5.2 Sources of Free Energy

We already know of one planetary environment in solar system history, suggested in meteorites, where liquid water and organics were present but in which little progress seems to have been made toward the origin of life (Chyba and McDonald 1995). Carbonaceous chondrite meteorites derive from parent bodies that may have seen liquid water for between ~10^4 and 10^8 yr, yet prebiotic chemistry in these meteorites seems not to have progressed very far beyond monomers such as individual amino acids and low concentrations of diglycine (Cronin 1976, Shimoyama and Ogasawara 2002). It is tempting to conclude that meteorite parent bodies simply lacked sufficient sources of free energy to drive substantial chemical disequilibria. This emphasizes the importance of assessing possible sources of free energy on Europa as a prelude to speculation about life on that world, however water- and organic-rich it might prove to be.

Discussion of energy sources for life on Europa began with Reynolds et al. (1983), who emphasized the difficulty of powering a substantial ecosystem by photosynthesis through Europa’s ice layer. Gaidos et al. (1999) extended this line of reasoning to argue that it is difficult to identify any sources of chemical disequilibrium on an ice-covered world lacking photosynthesis, with grim consequences for the prospects of life on Europa.

But if hydrothermal vents exist at the base of Europa’s ocean (Section 15.4.4), an ecosystem could be powered through methanogenesis or other reactions, for example, combining CO_2 with H_2 derived from fluid–rock reactions (McCollom 1999). Whether or not such hydrothermal activity is expected on Europa remains a matter of debate (Moore and Schubert 2000, McKinnon and ShocK 2001).

A second possibility is that life could take advantage of the disequilibrium production of O_2 and H_2 in the putative ocean due to the decay of ^{40}K (Chyba and Hand 2001). Recombination of these molecules by microorganisms could produce ~10^7 to 10^9 kg yr^{-1} of biomass today. Steady-state biomasses are difficult to calculate, because appropriate maintenance energies are difficult to estimate (Chyba and Phillips 2001). If we use a biological turnover time of ~10^8 yr, appropriate for Earth’s deep biosphere (Whitman et al. 1998), the above calculation suggests a potential biomass of ~10^{22} to 10^{23} kg, which could be compared with the terrestrial oceanic microbial biomass of nearly 10^{15} kg. This potential energy source would be largely independent of the thickness or heat loss mechanisms of Europa’s ice cover.

Two other sources of chemical disequilibrium for a eu­ropaon biosphere depend on the behavior of Europa’s ice shell. One is decay of ^{40}K in the ice shell; this mechanism is likely less important in the shell than in the ocean because K will be excluded from ice as water freezes (Chyba and Hand 2001). However, a second source is potentially important: the production of oxidants such as O_2 and hydrogen peroxide (Delitsky and Lane 1997, 1998, Gaidos et al. 1999, Chyba 2000a,b, Cooper et al. 2001) and organics such as formaldehyde (Chyba 2000a,b) in the uppermost meter of Europa’s
ice due to charged-particle bombardment followed by impact gardening to meter depths (Cooper et al. 2001, Chyba and Phillips 2001). This effect would be important for life on Europa only if the uppermost meters of Europa communicate with the putative ocean. If the timescale for this communication is 50 Myr – i.e., if Europa’s surface is mixed into the ocean once every 50 Myr on average – then ~10^7 to 10^11 kg yr\(^{-1}\) of biomass could be produced on Europa, depending on O\(_2\) production and microbial growth efficiencies (Chyba and Hand 2001). These various potential sources of chemical disequilibrium on Europa are summarized in Figure 15.19.

It has also been suggested that regions or niches might exist within Europa’s ice cover where photosynthesis may be possible within liquid-water environments (Reynolds et al. 1983). Some of these models rely on a cracks or melt-through events in a relatively thin ice layer (Greenberg et al. 2000, 2002a). Dissipative heating in Europa’s cracks may provide a different route to this end (Gaidos and Nimmo 2000). Were photosynthesis possible in limited environments on Europa, the energy available for life in these regions would swamp other sources. If life exists on Europa, and were it to have access to near-surface liquid-water environments, there would be strong selection pressure for it to colonize these environments because of this huge energy advantage.

15.5.3 Broadening the Definition of Habitability

The notion of “habitability” of a world once referred to those conditions suitable for human life (Dole 1964). The word has since come to refer to conditions less stringent: those necessary for stability of liquid water at a world’s surface (Kasting et al. 1993). The elucidation of Earth’s subsurface biosphere, and the discovery of a likely ocean on Europa, suggest that the universe may provide a much more expansive arena for life than the “classical” definition suggested (Sagan 1996, Chyba 1997, Chyba et al. 2000). But the only way to know whether life exists on Europa is to go and search.

15.6 SUMMARY AND CONCLUSIONS

Each of the Galilean satellites exhibits unique characteristics in terms of its surface evolution. Io (Chapter 14) is distinguished as a rocky moon and is the most volcanically active object in the solar system. Europa is primarily a rocky object but has an outer shell of water and a “lithosphere” of ice. Its surface history is characterized by tectonic deformation of the ice and local resurfacing by melted and/or ductile ice. The relative paucity of preserved impact craters on Europa suggests a rapid rate of resurfacing (Figueredo and Greeley 2004). Ganymede (Chapter 16), composed of about half water and half rocky material, shares some attributes with Europa by having extensive tectono-volcanic resurfacing of its ice-rich surface, but also has substantial terrains that preserve the record of cratering. Thus, the rate and intensity of internal activity is less than that of Europa. Finally, Callisto’s surface (Chapter 17) is dominated by impact craters of all sizes; yet, it also might contain a deep-seated “ocean” of liquid water. However, the high depth and low amount of internal energy have apparently precluded processes leading to endogenic resurfacing.

A key question for Europa remains unanswered. Does this moon currently have liquid water beneath its icy outer shell? Most investigators agree that many of the surface features described in the previous sections reflect the presence of liquid water (or at least warm, ductile ice) at the time of their formation, but the features do not necessarily reflect today’s conditions. However, given the relative youth of the features, it would seem unlikely that liquid water/warm ice would have frozen in the last ~60 million years, after a period of several billion years of its existence. This consideration, coupled with the results from the Galileo magnetometer suggesting the presence of a brine-rich ocean, leads most researchers to conclude that liquid water/warm ice is present today.

The question of Europa’s putative ocean is unlikely to be resolved definitively with currently available data. The experiments for the Galileo mission were conceived in the early 1970s (prior to the Voyager spacecraft flybys) and were not designed to address this question. Because of the high interest in Europa for astrobiological exploration, numerous formal (e.g., Space Studies Board 1999, 2003) and other studies (e.g., Chyba 1998) have considered the types of space flight projects that might be implemented. One sequence involves an orbiter around Europa that would (1) measure the degree of daily tidal flexing (which would enable the presence of an ocean to be detected and the thickness of the brittle ice to be constrained), (2) obtain global imaging at uniform resolutions and illumination conditions, and (3) map surface compositions. Should liquid water be detected, the next step would involve a landed system to make in situ measurements of compositions and conduct biologically relevant experiments. Subsequently, subsurface probes might be deployed to make observations and measurements below the radiation-intense zone on and near the surface.

The implementation of this sequence of missions was...
initiated with plans for the Europa Orbiter. Unfortunately, the escalating costs of this project led to its cancellation in 2002. Currently, various missions options are being reconsidered, including the Jupiter Icy Moons Orbiter, which would derive both power and propulsion from nuclear sources. The fundamental issue is whether to conduct the projects serially, as outlined above, or whether to combine elements into a larger project(s). For example, most groups agree that a lander of some type is required to address the key astrobiology questions for Europa. Thus, rather than fly an orbiter and lander separately, the two components might be combined.

In summary, the outstanding questions that need to be addressed for Europa include:

- Is there liquid water on Europa and, if so, what is its spatial distribution? If there is a globally distributed "ocean" of water, how thick is the layer of ice that covers it, and what are the properties of the liquid? If there is not liquid water today, was there any in the past, and what is the time dependence of its occurrence?
- Are the kilometer-scale ice rafts seen on Europa's surface a product of the movement of ice on an underlying liquid-water sea or through a warm, soft (but not necessarily melted) ice? What is the overall relationship between the surficial geologic units and the history of liquid water?
- What is the composition of the deep interior of Europa, including both the presumed silicate mantle and the iron or FeS-rich core? Is the core solid or liquid? Was there ever a global magnetic field produced by motions within the core? What are the dynamics of the interior, especially regarding the possible physical decoupling of the rotation of the surface ice from the deep interior and the possible non-synchronous rotation of the surface or of the entire satellite? What is the magnitude of tidal heating of the interior, and how is the heating distributed within the interior? Is the rock mantle convecting and is there active volcanism?
- What are the "absolute" ages of the features seen on Europa and what is the timescale for the key events in the evolution of the surface?
- What is the composition of the non-ice component (such as salts) of the surface materials that are seen in imaging and spectroscopic investigations? How do they vary over the surface? What are the source and history of these materials, and how do they relate to the geologic history of the surface and the potential for the at-least-intermittent presence of liquid water?
- What is the nature of the ice-tectonic processes that have affected the surface, and how are they reflected in the features that are seen (such as triple bands and spots)? Is there active or ongoing cryovolcanism?
- What is the composition of the neutral atmosphere and of the ionosphere? What are the sources and sinks of these species? What are the spatial and temporal variations in the atmosphere, and how do they relate to the physical processes that might control them? What is the composition of the magnetospheric ions that can sputter the surface, and of the sputtering products?
- What are the characteristics of the radiation environment at the surface of Europa (currently and in the past), and what are the implications for organic/biotic chemistry and the survival of life on the surface?
- What is the abundance of geochemical sources of energy that could support an origin of life on Europa or its continued existence? Is there extant life, or has there been life in the past? If there has been liquid water, access to biogenic elements, and a source of energy but there is no life present, what factors might explain the lack of occurrence of life, and does the potential exist for an independent origin of life in the future?

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