LIFE IN EXTREME ENVIRONMENTS
Life in Extreme Environments

by

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From our studies of some of the deepest parts of the oceans, the highest mountains, the coldest polar regions and the hottest and most arid deserts, we now know that life exists in numerous parts of the planet that, only a few years ago, were considered incapable of supporting life. In addition, data obtained from space research have provided tantalizing evidence of environments for possibly supporting life elsewhere in our Solar System.

Life in extreme environments is an appealing and exciting subject as it sits at a convergence point for two important questions for mankind: “How did life appear on our planet?” and “Is there life beyond our planet?” The harshness of Earth-based extreme environments offer the closest approximation to the conditions that probably existed when life first appeared on our planet but also offer potential analogues for conditions on other planetary bodies. Addressing the topic of life in extreme environments is also very relevant for one of today’s most crucial issues: the impact of human activity on ecosystems.

The investigation of life processes in extreme environments has a broad spectrum of relevance, including both societal and economical considerations. These exciting areas of research (whether considering microbes, plants or animals, including humans) are at an early stage and focus on environments that have in the past been difficult to investigate. However they are set to benefit tremendously from the new technological developments (e.g. robotics, information technologies, simulation techniques) as well as from the use of the rapidly developing tools of molecular biology and bioinformatics and are ideal targets for the consideration of species within their whole ecosystems. Actions should be taken to move research in this direction. At the European level the scientific community currently studying extreme environments is significant and well regarded but its structure is relatively fragmented. The benefits resulting from improved coordination and information exchange within this community are clear and implementing greater networking represents a significant challenge to Europe for the future.

In 2004, the European Science Foundation (ESF) launched a call for Expressions of Interest for research topics. The substantial response and high quality of ideas received as well as the broad diversity of domains covered encouraged ESF to set up an informal group of experts in November 2004 and to fund a large interdisciplinary
workshop in November 2005: the first integrated interdisciplinary initiative at the European level. It was aimed at identifying the points of view, priorities and recommendations of the European scientific community on these matters and the resulting report is to be published in the first quarter 2006. Alongside this initiative and complementing the report, the European Science Foundation is happy to have provided support to this special issue of Reviews in Environmental Science and Biotechnology on the topic of Life in Extreme Environments.

Professor Bertil Andersson  
Chief Executive  
European Science Foundation (ESF)
Access to glacial and subglacial environments in the Solar System by melting probe technology

Stephan Ulamec · Jens Biele · Oliver Funke · Marc Engelhardt

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Abstract A key aspect for understanding the biological and biochemical environment of subglacial waters, on Earth or other planets and moons in the Solar system, is the analysis of material embedded in or underneath icy layers on the surface. In particular the Antarctic lakes (most prominently Lake Vostok) but also the icy crust of Jupiter’s moon Europa or the polar caps of Mars require such investigation. One possible technique to penetrate thick ice layers with small and reliable probes is by melting, which does not require the heavy, complex and expensive equipment of a drilling rig. While melting probes have successfully been used for terrestrial applications e.g. in Antarctic ice, their performance in vacuum is different and theory needs confirmation by tests. Thus, a vacuum chamber has been used to perform a series of melting tests in cold (liquid nitrogen cooled) water ice samples. The feasibility of the method was demonstrated and the energy demand for a space mission could be estimated. Due to the high energy demand in case of extraterrestrial application (e.g. Europa or polar caps of Mars), only heating with radioactive isotopes seems feasible for reaching greater depths. The necessary power is driven by the desired penetration velocity (approximately linearly) and the dimensions of the probe (proportional to the cross section). In comparison to traditional drilling techniques the application of a melting probe for exploration of Antarctic lakes offers the advantage that biological contamination is minimized, since the Probe can be sterilized and the melting channel freezes immediately after the probe’s passage, inhibiting exchange with the surface layers and the atmosphere. In order to understand the physical and chemical nature of the ice layers, as well as for analysing the underlying water body, a melting probe needs to be equipped with a suite of scientific instruments that are capable of e.g. determining the chemical and isotopic composition of the embedded or dissolved materials.

Keywords Melting Probe · Subglacial · Europa · Mars · Antarctic Lakes · Ice · Technology · Life in extreme environments
Introduction

Ideas to use so-called “melting probes” for getting access to the subsurface layers of planetary ice sheets have been discussed for several years in connection with proposed Mars and Europa missions (Paige 1992; DiPippo et al. 1999; Carsey et al. 1999; Biele et al. 2002). More recently, some theoretical and experimental work has been done to understand the behaviour of such probes under extraterrestrial (very low temperatures, vacuum) conditions (Kömle et al. 2002, 2004; Treffer et al. 2006).

These investigations are driven by three major insights. First, there is a high scientific interest in exploring certain icy environments (Mars’ polar caps, Europa and other icy satellites) motivated by the search for traces of life in these extreme environments as well as interest in planetary climate history in the case of Mars. Second, robotic space missions with a mechanical drilling system for depths exceeding a few meters are mechanically very demanding. Third, contamination avoidance in respect to planetary protection requirements can be relatively easily fulfilled using melting probes, since the melting channel refreezes behind the probe and cuts off the contact to the surface. Moreover, decontamination of the probe is relatively easy to achieve with standard sterilization methods (Engelhardt 2006). Missions to explore the surface and sub-surface of icy satellites like Europa can use a wide variety of schemes and technologies. These differ in their cost and scientific return, with fly-bys and orbiters being relatively achievable, but yielding no in-situ knowledge of the subsurface. At the other extreme in terms of returned science and complexity is a device that is capable of melting its way through the satellite’s icy crust towards the putative ocean of water. Obviously, information about the biological activity within or under the Europan ice crust can be efficiently gained only with such an in-situ probe. Traces of indigenous biological activity, such as intact biomolecules, are unlikely to remain unaltered for long periods at the exposed surface of Europa (the radiation dose is \( \approx 1 \cdot 10^8 \) rad/month here. At greater depths, the radiation environment continues to decrease, reaching values comparable to the Earth’s biosphere below depths of 20 to 40 m (National Academy of Sciences 2000).

Table 1 summarizes parameters of icy surfaces of Mars, Europa and the Earth. Obviously, the environments require different technical solutions for melting probes. Dust content, salinity, temperature and radiation environment vary considerably between the three examples.

None of the existing melting probes is nearly mature enough for planetary applications without considerable technical development.

Melting probes

2.1 Past developments

The idea of a melting probe for the investigation of ice sheets can be retraced to the beginning of the 1960’s, when Shreve (1962) described a method for the calculation of the melt penetration efficiency of isothermal solid-nose hot points boring in temperate glacier ice. Shreve referred explicitly to Kasser’s experiments with a light ice-drill on glaciers (Kasser 1960). Philberth (1962) suggested a new method for measuring temperatures inside a glacial ice sheet. He designed a non-recoverable, instrumented melting probe, known thereafter as the Philberth probe, consisting of a cylindrical hull with an attached conically shaped head. The probe was connected to an external power supply by a cable (tether), unwinding from a coil integrated in the hull. For protection of the probe from damage caused by refreezing melt water after turning off the heating elements for ice temperature measurements, a significant part of the hull was filled with silicone oil of density \( >1 \text{ g/cm}^3 \) (Aamot 1967b).

The principal design of a cylindrical hull of small diameter attached to a conical, convex or even concave melting head is still characteristically for melting probes, because it is energetically favourable compared to other imaginable designs like e.g. a bowl- or egg-shape. The reason for this is that the melting probe’s penetration velocity \( v \) is inversely proportional to the cross-sectional
area of the probe, as described in more detail in the following section. Sufficient space for instrumentation can be obtained by increasing the length of a cylindrical probe, and therefore the length to diameter ratio of the first melting probes (including the Philberth probe) was about 25:1. A detailed description of the first Philberth probe can be found in Aamot (1967a).

Aamot (1970a) also reports four U.S. field experiments with early melting probes. In 1965, the first probe reached 90 m in Greenland ice before contact was lost. One-year later, the second probe ran smoothly for four days and reached a depth of 259 m after which it was stopped to observe temperature and pressure variations during the refreezing of the melt water. During the first day, this melting probe operated at 3380 W achieved a penetration velocity of 0.0755 cm/s (2.72 m/h). A detailed description of these first two field applications of melting probes, their instrumentation and the scientific results is given in Aamot (1967b).

During the 1968 EGIG expedition at Station Jarl Joset (Greenland), Philbert used two probes which reached depths of 230 and 1000 m and provided again more scientific and engineering information (Aamot 1970a). A new concept of a pendulum attitude stabilization for thermal probes was first introduced by Aamot (1967c, 1970b), with additional heating elements in the top part of a melting probe. In the 1980’s, the Polar Ice Coring Office (PICO) of the University of Nebraska constructed a similar probe with a telemetry link in its tether (Hansen and Kersten 1984). The PICO probe allowed constant temperature/ice flow measurements during the ice penetration phase as well as measurements of the melt water conductivity, and penetration depths between 100 m and 200 m were achieved. More recent work of the University of Nebraska on a probe for terrestrial applications is described by Kelty (1995).

In the 1990’s, the Alfred Wegener Institute for Polar- and Marine Research (AWI) constructed new probes for the investigation of Antarctic shelf ice. The AWI “SUSI II” probe (“SUSI” is the abbreviation for Sonde Under Shelf Ice) was successfully tested on the Rettenbach glacier (Sölden/Austria) in 1990 and achieved a depth of 60 m. The probe could be retrieved from the glacier because the melting hole did not refreeze within 8 h. Two-years later, during winter 1992/93, the probe was able to penetrate 225 m thick Antarctic shelf ice near the Neumayer station, and accessed the open water below the ice sheet (Tüg, H., personal communication 2002; Tüg 2003). The specifications of SUSI II were as follows: length 2.25 m, diameter 10 cm, power 3.4 kW (head), 600 V/8 A supply power, penetration velocity in the shelf ice: 2.93 m/h (220 m in 75 h), stabilization regulated over wire tension.

A further development of this probe, named SUSI III, was build for penetration of an ice thickness of 800 m, with length 3.6 m, diameter 14 cm, power 9 kW (head), 1000 V/12 A supply power. However, this probe failed because of a tether overheating in the winch compartment of the probe (AWI 1997). Another “routine” probe for lesser depths has been patented (Tüg 2003).

The melting probes described above have in common:

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<th>Table 1</th>
<th>Relevant parameters for melting probe application on Earth, Mars and Europa</th>
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<tr>
<td><strong>Earth (polar)</strong></td>
<td><strong>Mars (polar)</strong></td>
</tr>
<tr>
<td>Temperature [K]</td>
<td>220–270</td>
</tr>
<tr>
<td>Gravity [ms⁻²]</td>
<td>9.8</td>
</tr>
<tr>
<td>(atm.) pressure [mbar]</td>
<td>1000</td>
</tr>
<tr>
<td>Radiation environment [rad/day]</td>
<td>~ 0</td>
</tr>
<tr>
<td>Composition (of ice mantle)</td>
<td>H₂O, (some air)</td>
</tr>
<tr>
<td>Maximum pressure at bottom of ice layer (bar)</td>
<td>230–500</td>
</tr>
<tr>
<td>N.B.: phase transition of ice Ih occurs for p &gt;2000 bar.</td>
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• typical lengths from 2.2 m to 3.6 m,
• diameters of 10 cm to 14 cm,
• a total mass of hundreds kilograms,
• and a power usage of the order of 3 kW to 15 kW.
• penetration rates of the order of 2 m/h

Parallel to the technical further development of melting probes, a more sophisticated instrumentation of the probes took place. In particular, the SUSI probe of the AWI was instrumented for the investigation of microorganisms living inside micron-sized brine channels and pockets in sea ice and shelf ice. Parameters measured by this instrumentation cover oxygen and carbon dioxide content of the ice/water mixture, the amount of inorganic nutrients, and the pH-value (Tüg 2003). The PICO probe was upgraded to measure meltwater conductivity and micro-particulates.

With modern space flight and the advancing knowledge about the environmental conditions on e.g. Mars or on Jupiter’s icy moon Europa, the idea arose to use melting probes also for extraterrestrial applications. However, for the search for signs of life on Mars and especially on Jupiter’s moon Europa a new generation of melting probes on the basis of gathered experiences during the four decades of terrestrial applications is required. Critical parameters for any space mission are, e.g., the mass, the overall dimensions and the power supply.

Without doubt, the most advanced melting probe design is that of the Cryobot, initiated by the Jet Propulsion Laboratory in 1998 as an in-situ exploration and sample return vehicle for a future application on Europa (Zimmerman 2001). The Cryobot is allotted with integrated radioisotope thermoelectric generators (RTGs) providing 1 kW thermal of direct melt energy. The probe is 0.8 to 1 m long with a diameter of 12 cm and a mass of 20 kg to 25 kg in the flight version. Fluid thermal modelling and testing revealed that with 1 kW power, a melt rate of 0.3 m/h can be realized in very cold (100 K) ice; the water jacket maintained around the probe is 1–2 mm in width and the melt plume would not refreeze until 1.25 m behind the vehicle. An active heater system (warm water jet in front of the probe) allows faster penetration in dense ice. The Cryobot is designed as a fully autonomous robotic mole penetrator system for melting through an ice-pack of 3 km to 10 km thickness. Cardell et al. (2004) suggested a thermal probe for application also on the Martian polar caps for an investigation of Mars’ climate history, recorded in “polar layered deposits”.

2.2 Comparison with conventional drilling

Although this paper is summarizing technology and applications of melting probes one should mention the more conventional methods for ice penetration to outline benefits and drawback of the melting probe technology.

In particular ice-coring drilling (ICD), hot water drilling (HWD) and coiled tube drilling (CTD) need to be mentioned, since these technologies are well developed and have been successfully used in e.g. Antarctic ice. An excellent overview of these methods is given by Clow and Koci (2002).

When ice penetration is accomplished mechanically using rotary cutting and coring bits the drilling rate \( v \) for mechanical systems is given by

\[
v = \frac{eP_0}{AE}
\]

with drill-to-rock transmission efficiency \( e \), power output to drill \( P_0 \), hole cross section \( A \) and specific energy \( E \). According to Mauer (1968), \( E \) is of the order of 20 MJ/m\(^3\) for conventional drilling in “soft rock” (compressive strength up to 500 bar; ice has an compressive strength of 80–500 bar) and 50–170 MJ/m\(^3\) for oilfield rotary mining (roller bits, drag type) up to 2000 m depth. Mellor (1989) quotes values of 0.5 MJ/m\(^3\) to 5 MJ/m\(^3\) for ice, up to 14 MJ/m\(^3\) for frozen silt and 2 MJ/m\(^3\) to 6 MJ/m\(^3\) for frozen sand.

For cutting ice by a drill in thermal vacuum conditions resembling those expected, e.g., on Europa’s surface: a figure as low as 1.2 kJ per kg of cut ice has been quoted (DiPippo 1999).

These figures do, however, not include transmission losses and the energy for compacting of the cuttings and transportation to and discharge at the surface.
The specific energy for conventional drilling can be compared with the specific energy required for melting ice (>306 MJ/m³).

Ice core drilling (ICD) is an advanced technology, maximizing the recovery of ice samples and allowing geophysical logging methods to be applied. But it is also relatively slow and expensive to operate. Large quantities of drilling fluid are required. Application on Mars or Europa seems extremely challenging.

For drilling into ice in the 1 km range, relatively small hot water drilling (HWD)-devices can be used (Engelhardt et al. 1990). It is also possible to obtain short ice cores with this technology (Engelhardt et al. 2000. Since in case of HWD also the contamination is relatively low, a potential melting probe application on Earth needs to be carefully compared with HWD-technology for any advantages and disadvantages. HWD’s are very fast (i.e., drill rates on the order of tens of m/h are possible with the turbulent heat transfer at the ice interface), the energy expended for reheating is considerable (of the order of MW). Since hot (~80–90°C) water needs to be pumped into the borehole and due to the high power (fuel) demand of HWD, in case of an application on extraterrestrial bodies, a melting probe will have very clear advantages. There is no obvious solution for how to use HWD in vacuum.

Coil tubing technology (CTD) has first been used in 1991 (Sas-Jaworsky and Bell 1996) and undergone considerable development since. CTD’s are more compact and easier to operate than rotary drills. Since (in difference to HWD) material is drilled mechanically rather than using heat to melt the ice, they can be used also for deep (>4 km) drilling. CTDs are more and more commercially used (e.g. Gantt et al. 1998). A CTD for drilling into ice (CTDI) could be operated with various drilling fluids including hot water. CTD technology can also be used for drilling into bedrock, thus the problem of possible mineral deposits in the ice (which is a serious issue for melting probes) is obsolete. However, as discussed for HWD, the technology seems difficult for the use on Mars or Europa. A CTDI system, as proposed by Clow and Koci, (2002), capable of drilling a 3.5 km borehole through ice using ice coring, would weigh about 46 tons (including drilling fluid and fuel). Though being considerably lower in mass than a comparable system, based on HWD or ICD technology it is still not a trivial task to bring such a device into space.

2.3 Theory of melting probes

In a simple energy balance approximation, neglecting all losses, we can write the heat needed to progress a distance \( l \) in compact ice via melting as;

\[
\Delta W = Alq(c_p(T_F - T) + L_m)
\]

If the heating power is \( P \), then the melting speed \( v \) is

\[
v = lP/\Delta W
\]

\[
= \frac{P}{Aq(c_p(T_F - T) + L_m)}
\]

where \( A \) is the probe’s cross-section (m²), \( l \) is the probe length (m), \( c_p \) is the specific heat capacity of ice, \( \rho \) is the ice density, \( L_m \) is the melting enthalpy of ice. More generally, it is the energy needed to change the phase (solid–liquid or solid–gaseous), \( T_F \) is the melting temperature of ice (K), \( T \) is the local ice temperature (K).

Precise numerical values and correlation equations for the physical properties of ices are given in the appendix.

Note that \( \rho \) and especially \( c_p \), are temperature dependent; for low temperatures \( T \), one has to use averages over the temperature range \([T; T_F]\). Also, \( T_F \) depends (weakly) on ambient pressure and on salt content (freezing point depression).

The most important point here is that the melting velocity scales with the inverse of the cross-sectional area of the probe. Hence the usual design is a cylindrical tube with a large (>10) aspect ratio (length/diameter) to obtain a volume sufficient for subsystem, payload and tether integration.

Equation (2) gives only a rough estimate of the penetration velocity, in fact, it relates the minimum power requirement \( P_0 \) to a given melting velocity. In fact, insufficient heat causes the probe to freeze in i.e. to stall; excessive heat
produces an oversize hole and wastes power by raising the meltwater’s temperature far above the phase transition temperature. To obtain a more accurate result loss factors need to be analysed: most importantly, losses due to lateral conduction in the ice and melting a slightly bigger hole than the cross section of the probe (depending on the melting speed itself, the applied power and the surface of the probe); further, friction losses between the ice wall of the melted channel and the surface of the probe, due to the viscosity of the thin water layer located in between or dust mixed with the ice; radiative losses; and losses in the tether.

The lateral conduction losses have been estimated by Aamot (1967c) for a cylindrical probe with the constraint that the melt water must just stay liquid all around the probe’s hull. The integrated lateral power requirement is given by:

\[ P_{\text{cond}} = \frac{4\lambda T}{R_{\text{probe}} \pi^2} \int_0^{L_{\text{probe}}} \int_0^\infty e^{-\kappa u^2 s/v} \frac{u J_0(R_{\text{probe}} u) + Y_0(R_{\text{probe}} u)}{u^2} du ds \]  

(3)

where \( u \) is the integration argument for the Bessel functions \( J_0 \) and \( Y_0 \), and \( s \) the spatial coordinate along the length \( L_{\text{probe}} \) of the probe. The thermal conductivity of ice is represented by the heat diffusion coefficient \( \kappa = \lambda / (\rho c_p) \) with the thermal conductivity of ice, \( \lambda \).

This (numerically difficult) integral can be conveniently approximated by (Aamot 1967c, and appendix A):

\[ \frac{P_{\text{cond}}}{R_{\text{probe}}^2 v (T_F - T)} = a x^b \]  

with \( x = L / (v R_{\text{probe}}^2) \)

\[ = t^* / R_{\text{probe}}^2 \]  

in \( \text{s/m}^2 \),

valid for \( 5 \cdot 10^4 < x < 10^8 [\text{s/m}^2] \)

with the fit constants \( a = 932 \text{ Ws/K/m}^3 \) and \( b = 0.726; t^* \) is called “characteristic time”. (4a,b)

\[ P = P_0(A, v) + P_{\text{cond}}(A, L, v) \]

Hence the total power required by the probe to penetrate the ice with a velocity \( v \) is given by \( P \).
Note that a correction has been applied for the calculation Figs. 1 and 2, since the data in Aamot (1967c) are calculated for water ice near 273 K, while the thermal conductivity ($\propto 1/T$) and specific heat capacity ($\propto T$) vary considerably over an extended temperature range. This has been considered (compare Eq. 3) by multiplying $T^*$ with $(2T_F/T_F + T)^2$ and the constant $a$ with $(2T_F/T_F + T)^2$ which has the effect to transform the heat diffusion coefficient $k$ and the heat conductivity $\lambda$ to their values at the mean temperature $(T + T_F)/2$.

While, e.g., the penetration rates given in (Zimmermann 2001) in $-10^\circ$C ice can be reproduced within 17% (the measured value is lower, probably owing to other losses than lateral heat conduction), the model becomes uncertain at very low ice temperatures due to the approximations involved. It is an ongoing effort (Kaufmann, E., private communication 2006) to re-calculate the lateral heat conduction integrals (Eq. 3) with proper thermophysical quantities over a wide temperature range and develop numerical thermo-mathematical models for melting probes in conjunction with verifications in extremely cold ice.

Aamot’s theory also permits to calculate the rate of lateral heat requirements along the length of the probe, which is important for design and heating regulation.

A theory describing the flow of melt water around a hot point has been developed by (Shreve 1962), but only for the case of isothermal ice at or very slightly below the melting point (“temperate ice”) under terrestrial pressure conditions and non-turbulent flow – lateral heat conduction into the surrounding ice is entirely neglected. Interestingly, the efficiency ($P_l/P_o$) of the hot point then depends mainly on the ratio $PS/RW^{1/4}$ with $P$ heating power, $S$ shape factor of frontal surface (e.g., 1 for a flat surface, around 0.6 for the usual blunt paraboloids or half-spheres), $R$ radius, and $W$ weight (corrected for buoyancy) of the probe.

As Kömle et al. (2002, 2004) pointed out, in an atmosphere-less environment (such as Europa), when the probe is not fully immersed in solid ice, the ice might not melt but sublime. The latent heat of sublimation $L_s$ at 270 K is $\sim 8$ times higher (2836 kJ/kg/K at $0^\circ$C) than the latent heat of fusion, so initial progress of the probe at the surface will be accordingly slower until it has been covered by solid ice. Then, melting can occur once a gas pressure greater than the triple point pressure of 611.6 Pa can be sustained around the probe. Tests have been performed on this issue in thermal vacuum chambers (Treffer et al. 2006 and see Section 2.4). The result is that in solid ice, sublimation obviously dominates only the first few cm until the head of the probe has penetrated the ice. An “ice collar” forms very quickly around the rim and obviously suffices to raise the pressure in the tip zone over the triple point, such that very soon melting dominates the penetration process. The situation seems to be different, however, in porous ice where the sublimated gas either escapes or recondenses in the open pores of the surrounding material (Kömle et al. 2004). Here, sublimation dominates at least until the melting channel is completely closed behind the probe, i.e. at least to a depth that is equal or greater than the probe’s length.

It is also worth mentioning that all thermal balance considerations imply a close contact of the melting probe’s hot nose with the ice/meltwater. If, by obstacles or blocking, a void develops between the hot point and the ice, heat is transported predominantly by radiation, the hot point’s temperature increases rapidly and the melting speed drops until the contact is re-established.

### 2.3.1 Other losses

**Viscous friction.** The Reynolds number applicable for the melt water flow is

$$Re = \frac{\rho_w v h}{\mu_0}$$

with water density $\rho_w$, water viscosity (at $-10^\circ$C) $\mu_0$, width of melt water jacket $h$ and melt speed $v$. Typical values ($v = 10^{-3}$ m/s, $h = 1$ mm, $\mu_0 = 1.8 \cdot 10^{-3}$ Pa s) yield Reynold numbers of the order of 1, while the transition to turbulent flow occurs at $Re > 1160$. Thus, laminar flow is assured and the viscous friction can be estimated as
\[
F_{\text{visc}} = \eta A \frac{dv}{dx}, \quad \text{with} \quad A \simeq 2\pi (R^2 + RL),
\]

giving very small numbers (for typical melting probes on Earth, \(F_{\text{visc}}/F_{\text{gravity}} \approx 1 \times 10^{-6}\)). Since the viscous friction is so small, it can be neglected even for low-gravity environments like on Europa.

**Tether losses.** If a tether is payed out by the probe, it will conduct heat from the warm coil towards the melt water and cold refrozen ice behind the vehicle. However, the temperature gradient is not very steep as the tether is surrounded by \(\sim 0^\circ C\) meltwater for a considerable (order of dm) length. With a typical copper cross section (1 mm\(^2\)) of a power line tether, a gradient of (300–380) K along 1 m, the thermal conductivity of copper, 400 W/mK and a heat transition coefficient 350 W/mK we estimate a thermal loss of 0.04 W, which is negligible.

**Radiative losses.** For a typical melting probe advancing in very cold (80 K) ice, of the order of 10% of the heating power are radiated as mid-IR radiation (Stefan-Boltzmann law with emissivities of probe and ice both set to 1). However, practically all radiation is immediately absorbed in the melt water jacket:

- 98% of a black body’s radiation energy is emitted between 4.8\(\mu\)m and 76\(\mu\)m
- The absorption coefficient of water in this wavelength range is >100 cm\(^{-1}\), mostly around 1000 cm\(^{-1}\) (Irvine and Pollack 1968).
- Thus, the total absorption in a 0.1 mm layer of water is >99.99%

We conclude that the thermal radiation remains in the melt budget and practically no losses occur.

**Overheating losses.** Although counter-intuitive, the addition of more heat does not necessarily increase the melt rate proportionally: since water is a good insulator (0.56 W/K/m compared to 2 W/K/m for ice at 273 K), raising the temperature of the melt water significantly above freezing results in a widening of the melt water jacket around the probe and a subsequent decrease of the heat transfer to the actual ice interface. At very high power densities (>3.25 MW/m\(^2\)), film boiling further decreases the heat transfer. Thus, the most efficient way to create the phase change is to input just enough heat to initiate melting (Zimmermann 2001).

**Solid friction.** By this we mean friction of the probe surfaces with solid (dust) admixtures. It is extremely difficult to estimate or model and can obviously lead to blocking (see also Section 2.4.7.).

In the following we write the sinking speed as follows, to include the case of porous ice, mixture of sublimation and melting, and conductive losses:

\[
v = \frac{P}{A\rho(1 - \Pi)\bar{E}(\bar{c}_p(T_F - T) + L_{\text{eff}})}
\]

where we have introduced; \(\Pi\): the porosity of the ice (in Greenland firn ice, typically 40% at 15 m depth, 15% at 50 m and \(\approx 0\) for depths exceeding 100 m); \(E\): the loss factor, \(E = 1 + P_{\text{cond}}/P_0\), \(L_{\text{eff}}\): the effective heat of phase change, i.e. a value between \(L_m\) and \(L_s\) depending on the actually dominant process; \(\bar{c}_p\): mean heat capacity between \(T\) and \(T_F\); \(\bar{\rho}\): mean solid ice density between \(T\) and \(T_F\).

Equation (4) has been compared to the experimental results of Kömle et al. (2002) covering both pure melting and sublimation in compact and porous ice (snow): the result is that with their probe design (sphere of 4 cm diameter) and power levels (25 or. 60W, respectively), \(E = 2.0 \pm 0.1\) as predicted. for an ice temperature of 80 K.

2.3.2 Ices other than water

It is interesting to compare the performance of melting probes in water ice with its performance in other cryogenic solids, notably the astrophysically interesting ices CO\(_2\), CO, and CH\(_4\). Table 2 gives an overview over the thermophysical properties (Phase change temperature, density, specific heat capacity, phase change enthalpy) of these ices, compared with water ice Ih; the last row gives the ratio of melting velocity to the one in water ice, all.
other conditions being equal (and $T_{\text{ice}} = 1/2 T_F$). As expected, the melting process in carbon monoxide and methane is about five times more efficient than in water ice; in carbon dioxide, however, it reaches only 40% of the performance in water ice because (down to a depth of about 1300 m on Mars, at least) the phase change is normally by sublimation due to the very high triple point pressure of carbon dioxide, ~74 bar.

2.4 Main technological issues

2.4.1 Gravity dependence

Gravity exerts a second-order effect on the penetration of melting probes. The efficiency of the idealized hot point is weakly dependent on the contact pressure of the melt surface on the ice (Shreve 1962). The force of gravity is of course necessary to set the direction of penetration, but does not (to first order) control the melting velocity. Only the possibly non-negligible friction between and sediments needs to be overcome by the weight of the probe. Thus, melting probes would not function under microgravity environments. However, gravity on Europa and Mars is 1.31 m/s$^2$ and 3.73 m/s$^2$, respectively, and not considered a major issue.

2.4.2 Prevention of blocking

Great care has to be taken, particularly in cold ice, to ensure that the re-freezing of ice laterally and behind the probe is slow compared to the melting velocity, in order not to jam the probe. While this is not an issue for e.g. terrestrial glaciers ($t \geq -10^\circ\text{C}$), it is critical for very cold ice as experiments at 77 K have shown (Treffen et al. 2006). Proposals for low temperature melting probes foresee heating elements distributed along the length of the probe and careful heating control for that reason.

The issue of blocking due to contaminants in the ice is discussed in Section 2.4.7.

2.4.3 Attitude control

A melting probe requires a means of attitude control if its vertical attitude is to be maintained. Such probes are inherently top heavy because they stand on their tip – the hot point. The slightest deviation from the vertical results in an increasing tendency of the probe to “lean” and “topple over”. This is almost inevitable in inhomogeneous ice (with voids, cracks, sediment layers etc.). This instability must be counteracted or eliminated. Mercury steering and “pendulum steering” are described in the literature (Aamot 1967a), with mixed results. The successful AWI “SUSI II” probe (Tüg, H., personal communication 2002; Tüg 2003) used a simple mechanism based on information of the hot nose temperature and the tilt of the probe to control the friction clutch of the probe’s tether/cable pay-out mechanism.

Active attitude control can also be based on controlling the heating power (e.g., laterally in the tip, between tip and upper annular heated ring, etc.). In a RHU-heated probe (see Section 2.3.5 below), the required heat distribution could be controlled by means of heat pipes with mechanical valves.

More elaborate techniques for active steering of the probe around obstacles, involving an acoustic obstacle detection sonar and a steerable probe pushing device, have been proposed by DiPippo et al. (1999).

2.4.4 Pressure

The usual approximation of the ambient pressure $p$ in bulk ice (valid if the ice is in a secular hydrostatic equilibrium) is

<table>
<thead>
<tr>
<th></th>
<th>CO$_2$</th>
<th>CO</th>
<th>CH$_4$</th>
<th>H$_2$O</th>
</tr>
</thead>
<tbody>
<tr>
<td>$T_{\text{f}},$ melting temperature (°C)</td>
<td>216.6</td>
<td>68.1</td>
<td>90.7</td>
<td>273.1</td>
</tr>
<tr>
<td>$c_p,$ specific heat capacity (kJ/kg°C)</td>
<td>1.38</td>
<td>1.90</td>
<td>2.35</td>
<td>1.13</td>
</tr>
<tr>
<td>$L_v,$ phase transition enthalpy (kJ/kg)</td>
<td>573.3</td>
<td>29.9</td>
<td>58.6</td>
<td>333.4</td>
</tr>
<tr>
<td>$\rho,$ density (kg/m$^3$)</td>
<td>1540</td>
<td>920</td>
<td>500</td>
<td>920</td>
</tr>
<tr>
<td>Penetration velocity relative to water ice, $T_{\text{ice}}=1/2 T_F,$ losses regarded as equal</td>
<td>0.40</td>
<td>5.1</td>
<td>5.4</td>
<td>1</td>
</tr>
</tbody>
</table>
\[ p = \rho g h \] where \( \rho \) is the ice density and \( g \) the local gravity, \( g = 9.82 \text{ m/s}^2 \) on Earth, \( g = 3.73 \text{ m/s}^2 \) on Mars and \( g = 1.31 \text{ m/s}^2 \) on Europa.

We get \( p \approx 400 \text{ bar} \) in 30 km depth on Europa as well as in 4000 m depth for Lake Vostok (see also Table 2). The probe must not only survive this pressure but, more importantly, the pressure transient when reaching the ice/water interface. This pressure jump should not occur if ice and water are both in hydrostatic equilibrium, but has been observed on Earth (Tüg, H., personal communication 2002). Aamot (1968) also reported measurements of pressure jumps during stopping or restarting the probe (freezing or melting of surrounding material with associated volume and pressure changes) up to 88 bar.

Two basic system architectures with respect to operational pressure may be considered: First, a “dry” system based on a pressure tight vessel, capable of withstanding structurally the maximum external pressures and housing all system components with sealed windows and openings for e.g., instrument access to the environment and tether payout. Second, a “wet” architecture where at least parts of the probe can either be flooded with meltwater, or are immersed in e.g., silicone oil. The latter concept has been used, for example, in the Philberth and AWI probes (for the tether storage canister section of the probe).

### 2.4.5 Power supply

The traditional power supply design is by cable, paid out from a coil stored in the aft of the melting probe. Delivering heating power via a cable to great depths is challenging due to both mechanical and electrical constraints, but simplifies communication and steering (attitude) issues.

Typically, high voltages (1000 V, DC) are used for the power supply to minimize losses in the tether and housekeeping and science telemetry is multiplexed on the DC supply voltage. Thus, two wires, of which only one has to be insulated, suffice for power and telemetry. Additionally, it is preferable to include a mylar tether that bears the weight of the probe and is used for steering.

Clearly the thickness of the cable, along with the available total volume for the storage of the coils, is the limiting factor for the maximum possible penetration depth. For the AWI’s SUSI probes, a maximum depth of the order of 1000–2000 m (for an optimized design) has been estimated (Tüg, H., personal communication 2002). However, concepts for Philberth probes described in (Aamot 1968) claim a feasible depth of 3000 m or more.

The use of power intensive devices such as ice-melting probes in the outer Solar System strongly points to radioactive units for heating and power supply. The traditional space application RHU (Radioactive Heater Unit) technology is based on \(^{238}\text{Pu}\) (specific thermal power 0.4 kW/kg, half-life 88 years). For an Antarctic application \(^{45}\text{Ca}\) seems to be an attractive alternative since it is not explicitly excluded by the Antarctic Treaty and is a beta radiator (half life: 162 d) so no gamma shielding is necessary. Its decay produces about 40 kW kg\(^{-1}\) and its decay product is \(^{45}\text{Sc}\) which is stable and non-toxic. The availability of \(^{45}\text{Ca}\), bearing in mind the short half-life, needs to be investigated or other isotopes with similar properties need to be found.

In conclusion, a depth of 4000 m (Lake Vostok) can apparently only be reached by either RHU power supplies or by launching the probe from the bottom of the already existing borehole just a few 100 meters above the ice/water interface. For small Mars polar cap probes, power supply by cable from a solar generator at the surface (during polar summer) seems to be possible (a heating power of the order of 25–50 W seems to be sufficient for small probes, see Kömle et al. 2002), while for Europa only heating by RHU’s appears feasible.

### 2.4.6 Communications

The impracticality of autonomously guiding the probe back to the surface with any accuracy means that a melting probe in a Europan setting will, at first, be a one-way mission. (Sample return missions have been suggested by Biele et al. (2002). However, experimental proof of the concept is lacking at the time being and it is not
foreseeable how close to the surface module the returned probe would re-appear). Also the strategy to recoil the tether, and thus, “climbing” back to the surface may be considered. A problem here might be the need to heat the probe at the top (i.e. close to the tether/cable coil).

Information will therefore have to be conveyed to Earth by a communication system that uses several separate links. Terrestrial projects can employ cables to transmit power and communication data. Although such tethered systems have been examined for planetary ice-probes, they are likely to be limited to relatively shallow exploration depths of the order of a few kilometre.

The problems of communicating between a Europa lander and the Earth are relatively trivial when compared to the problem of sending data through many kilometres of impure water ice. However, an advantage of using radioactive sources for heating and power-raising in a melting-probe is that the vehicle is then relatively ‘power-rich’ and radio communication methods may be viable between a probe and a lander.

Unfortunately, the presence of salts in water ice can radically increase the attenuation experienced by microwave signals, to the extent that studies have considered deploying relay transceiver pods behind the probe as it melts through Europa’s surface. Typical propagation figures indicate that microwave links could pass 10 kbit/s over a few hundred metres with powers of around 200 mW. Realistically, lower rates will be achieved when the effect of adding salts and scattering bodies such as meteoritic debris are considered. Other methods of avoiding the high losses faced by high-frequency signals could involve using much longer wavelengths, which would yield lower bit-rates, and would require a mass memory for telemetry buffering.

Acoustic waves can be transmitted through layers of liquid and solid water; the achievable data rate would be very low and strongly influenced by the unknown characteristics of the ice (density, porosity, layering, mechanical discontinuities, varying attenuation, multiple paths, ambient noise, thermal cracking, reverberation etc.). Therefore, at the moment the use of a tether line (including a coaxial cable and/or optical fibers) preliminarily appear to be the most reliable, or only, solution, in spite of the great line length required, the need of a storage canister and a technique to manage its safe deployment. It could be based, however, on the mature technologies that exist in wire guided missiles, sub sea torpedoes and remotely operated submersibles, with the necessary improvements due to the particular environment and operating conditions concerned.

2.4.7 Ice contaminants

One major problem regarding the melting into natural ices is intrusion of components that cannot be molten, like dust or salts.

So far, no satisfactory solution has been demonstrated (nor do empirical data exist to our knowledge) to melt into dusty ice, since the dust concentrates underneath the melting tip, building up an insulating layer and increasing the friction. Mechanisms in addition to “pure melting” need to be applied. (di Pippo, 1999). In context with the JPL Cryobot an active water jet system is proposed, splashing/pumping away the debris accumulating at the melt head (Zimmermann 2001).

We would like to emphasize methods combining penetration techniques of melting with those of a “mole” (Gromov 1997). Such a device which can hammer itself e.g. into regolith (in fact, any compressible/porous medium) has a high degree of maturity and was part of the payload of the Beagle 2 lander on Mars (Richter et al. 2001, 2004). A combined melting-hammering shall allow penetration into dusty ices. However, development of a breadboard system combining these two concepts is still to be done.

Regarding salts in the ice one needs to differentiate according to their solubility. For non soluble layers similar solutions as for dust should be applied. Soluble salts like NaCl will be dissolved in the melt water but lead to a higher concentration in the melt water plume lagging the probe, compared to the re-frozen ice. This, however, will not lead to any blocking of the
 probe, even if some salt may precipitate behind the vehicle.

2.5 Breadboard tests for planetary applications

A number of experiments have been performed at DLR in a cold lab as well as in a vacuum chamber. A modified melting probe based on a prototype provided by the Alfred Wegener Institut für Polar- und Meeresforschung (AWI) was used for these experiments (Treffer et al. 2006).

The main part of the probe consists of a copper hemisphere melting head (11.5 cm in diameter), containing three 230 V heating elements with a power of 200 W, each. The heating elements can be operated independently from each other but usually the full heating power of 600 W (measured: 580–645 W depending on the AC voltage) was employed.

An upright aluminium frame was used for vertical suspension of the melting probe. Implementation in the vacuum chamber allowed a centred positioning of the probe above the ice container. The probe was attached with a wire to a strain gauge used to determine if the probe is in contact with the ice surface in order to control the tether mechanism (see Figure 3).

Freeze-in could not be isolated from zero melt-head contact pressure. Under nominal operation, the contact pressure changed by about one order of magnitude between “sitting on the ice” and “hanging in the melt hole surrounded by melt water”; an empirically determined force limit (that included the interval between zero and full immersion in melt water and, consequently, a range in buoyancy) was used in a control loop to feed more tether in an automatic way.

2.5.1 Experimental results

The experiments performed in both, cold lab (atmospheric pressure, solid ice at 243 K) and vacuum chamber ($p < 1$ hPa, solid ice at $-100$ K) demonstrated very well the applicability of the theory, as mentioned above. Melting velocities in vacuum were consistent with the predictions according to formulae (4, Aamot-fit etc.): $L = L_m$, efficiency $E \approx 23\%$ for $T_i = 100$K, melt velocities $(2.5...3.7) \times 10^{-5}$ m/s, effective length 0.2 m due to only partial penetration.

One important result is the fact that, in compact ice, the melting channel freezes again very rapidly, so the probe’s hot point is in an almost closed cavity already shortly after penetration, increasing the pressure beyond the triple point pressure. Thus, liquid water can exist, leading to faster progression (by melting, not sublimation).

The drawback of this effective re-freezing is that the probe itself got stuck at the unheated aluminium tube in the low temperature vacuum experiments. Clearly, the lateral heat requirements were not fulfilled. Thus, future
experiments will need controlled heating over the complete length of the probe.

3 Fields for future application

3.1 Terrestrial applications (Antarctica)

Deep underneath the Antarctic ice sheet there are tremendous concealed lakes of water kept in the liquid state by the enormous pressure of the overlying ice on the one hand and the continuous heat flux from the Earth's interior on the other hand. These lakes have been isolated from the atmosphere for several hundred thousand up to one million years. Until today about 145 of such lakes are detected (Siegert 2000; Siegert et al. 2005).

The lakes are of enormous scientific interest as a potential and unique habitat for life, having developed in a secluded environment. The environment in these habitats is considered to be similar to the conditions as they may exist also in extraterrestrial cases, e.g. on Europa (Priscu and Christner 2004).

The largest and one of the most interesting lakes is Lake Vostok, named after the Russian station located above the outer rim of the lake. The huge ice-sealed underground Lake Vostok lies four kilometres below the surface of the central Antarctic ice sheet. The ice sheet ranges in thickness from 3,800 to 4,200 meters. The lake is 200 km long by 40 km wide (at its widest point) and is divided into two deep basins by a ridge (Nadis 1999).

Figure 4 shows a radar image of Lake Vostok.

As a result of the special and isolated environment of the lake it can be expected that original life forms developed in an unprecedented way (Siegert et al. 2001).

Because of this particularity any investigation of the lake needs to guarantee minimum contamination with both bacterial life from the surface and any potentially toxic material.

Thus, conventional drilling appears problematic. It is necessary to use methods for investigation that satisfy the strict requirements of planetary protection, as applied e.g. for the exploration of Mars, and foreseen for any future Europa mission.

An international drilling project to Lake Vostok that started in the 1980s was suspended after growing awareness of this problem. The drilling was stopped at the depth of 3,623 m, or approximately 120 m above the surface of the lake (Giles 2004). A further 27 m of core was drilled in 2005/6 and lake penetration is now scheduled for 2007/2008 if the contamination problem has been solved then.

A sterilized melting probe, penetrating through a progressing cavity is an adequate solution to investigate the water of the lake in situ with minimized risk of contamination (Treffer et al. 2006).

3.2 Jovian Satellite Europa

One of the most fascinating bodies in the Solar System from an exobiological point of view is Europa, one of the Galilean Satellites of Jupiter. Table 3 lists the main physical and orbital parameters of Europa.

As the images provided by the Galileo Probe have revealed its surface is covered by a very young layer of water ice. The surface structures (showing almost no craters) indicate recent
movement of the ice crust and allow speculations about a global sub glacial ocean (Greenberg and Geissler 2002). An overview of the Galileo mission is given by Harland (2000).

Indications of water ice on the surface by infrared astronomy were already published by Kuiper in 1957 and later confirmed by Pilcher et al. (1972). During the Voyager 1 and 2 missions, when the two spacecraft reached the Jovian system in March and July 1979, respectively, images were taken that showed various surface structures, leading to the nomenclature of the observable features. An exhaustive overview of the results of the Voyager mission is given by Lucchitta and Soderblom (1982).

First considerations of a possible liquid ocean underneath the ice crust were given by Cassen et al. (1979) and after Voyager by Reynolds et al. and Squyres et al. both in 1983.

More recent papers include the Galileo results for a more detailed analysis of the satellite’s internal structure (Anderson et al. 1998; Khurana et al. 1998).

The surface temperature of Europa is about 100 K. The internal heating, leading to the (probable, but not proven) liquid ocean is explained by tidal heating and (to a much smaller extent) radioactive decay in the interior.

The effects of tidal heating of a satellite around Jupiter can most impressively be seen in the violent active volcanism of Io, the innermost of the Galilean satellites, were enormous eruptions were first detected by the Voyager spacecraft (Morabito et al. 1979). This was predicted by Peale et al. in 1979, when realizing that Io’s orbit has a substantial eccentricity. Shortly after the Voyager flybys the heat dissipation was recalculated (Cassen et al. 1980).

Figure 5 shows an eruption, imaged by the Galileo spacecraft in 1997.

Extrapolating the tidal heating rate of Io \(10^{14} \text{ W}\) to Europa, a value of about \(6 \times 10^{12} \text{ W}\) can be estimated (Greenberg 2005). A more elaborate model leads to higher values of \(9 \times 10^{12} \text{ W}\) internal heat dissipation or \(0.3 \text{ W/m}^2\) (O’Brien et al. 2000).

There is no generally accepted model for the thickness of the ice crust available yet. Values vary between >50 km and only a few kilometres (Cassen et al. 1979; Squyres et al. 1983; Ross et al. 1987; Ojakangas and Stevenson 1989). In addition there exists a model explaining Europa’s surface and magnetic field features with an underlying sheath of viscous icy material (and no liquid water at all) (Pappalardo et al. 1998).

---

**Table 3** Europa’s physical and orbital parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Equatorial radius [km]</td>
<td>1569</td>
</tr>
<tr>
<td>Mass [kg]</td>
<td>(4.8 \times 10^{22})</td>
</tr>
<tr>
<td>Surface gravity [m/s²]</td>
<td>1.31</td>
</tr>
<tr>
<td>Mean density [g/cm³]</td>
<td>3.01</td>
</tr>
<tr>
<td>Escape velocity [km/s]</td>
<td>2.02</td>
</tr>
<tr>
<td>Mean orbital radius [km]</td>
<td>671,100</td>
</tr>
<tr>
<td>Orbital period [d]</td>
<td>3.551</td>
</tr>
<tr>
<td>eccentricity</td>
<td>0.009</td>
</tr>
<tr>
<td>inclination</td>
<td>0.47°</td>
</tr>
</tbody>
</table>

---

**Fig. 5** Volcanic eruption on Io; the plume on the limb (Pillan Patera) is 140 km high. Image taken by Galileo, June 1997 (NASA/JPL 2005)
The idea of habitable zones on Europa has been discussed by Reynolds et al. (1983) and later by Greenberg et al. (2001); a very comprehensive overview of the physics of this satellite is given by Greenberg (2005).

Figure 7 shows the trailing hemisphere of Europa, as imaged by the Galileo camera (September 7th, 1996) in natural (left) and enhanced colours. The very low number of impact craters is a clear evidence for the young age of the surface terrain and, thus, of ongoing restructuring of the ice crust.

Figures 8 and 9 show surface features which are a strong indication for liquid underneath a thin crust (NASA/JPL URL).

An intrinsic problem of the investigation of Europa’s ice crust and its putative ocean is the penetration of the ice with a device suitable for a planetary exploration mission. Drill rigs appear far too heavy and complex to be implemented into currently considered Lander-missions to the Jovian system.

A melting probe, powered with radioactive heater units (RHU’s) seems to be a relatively low-cost, low-mass option to reveal some of the information, hidden underneath thick ice-layers.

This has been proposed e.g. by DiPippo et al. (1999) and Ulamec et al. (2005).

Even if the average thickness of Europa’s ice crust may be high, and thus, the liquid water difficult to reach, there may be areas where the ice is much thinner (only a few tens of meters), e.g. at the ridges, where melt-through may occur.

Figure 6 shows two possible scenarios of the internal structure of Europa (NASA/JPL 2005).

The most widely accepted model of the interior of Europa assumes a liquid ocean, warmed up by tidal forces. The idea, to find on the bottom of this ocean an environment comparable to black smoker environment at the bottom of the terrestrial oceans, can be seriously considered. Of course, this is speculative, but it is a fascinating area to be explored!
In such areas, melting through the thin ice may very well be possible, in an acceptably short time. At the same time, rather fresh material from the global ocean might be found at or close to the surface at these places (Greenberg, R, personal communication 2005). Once a melting probe is below the destructive radiation of the surface layers (~10–40 m), frozen ocean material might be sampled without actually accessing open water.

3.3 Mars polar caps

In 1666, the French astronomer and mathematician Giovanni Domenico Cassini observed the bright polar ice caps of Mars, visible already through a standard telescope. Figure 10 shows an image taken with the Hubble Space Telescope. The question, whether or not the Martian polar caps are consisting of water ice remained unanswered until the dawn of the age of modern space flight. The existence of the famous channels observed by Schiaparelli in 1877 and some other astronomers could be
definitely negated, when Mariner 4 as the first spacecraft sent 22 pictures of the Martian surface back to Earth during its flyby at the planet in July 1965. On the pictures various craters similar to that found on the moon were visible, and some of them appeared to be covered with frost (Godwin 2000). More pictures were obtained by the following flybys of other spacecraft, and in 1976 first compositional data of the Martian soil were transmitted to Earth by the Viking 1 and Viking 2 landers. The two Viking probes were the first to remain intact on the surface of Mars and collected valuable data, e.g. about the Martian weather. Although both landers were equipped with instruments for the detection of microbial life in the Martian soil, no evidence for present or past life was found (Ballou et al. 1978). However, the interesting places on Mars where microbial life forms could exist have not been explored until today. These places should contain water and should shelter microorganisms from the solar UV radiation that hits the Martian surface unfiltered.

The average temperature on the surface of Mars is – 63°C, with a maximum temperature of +20°C at the equator during summer, and a minimum of – 140°C at the poles in winter. The Martian atmosphere is rather thin with a surface level pressure of roughly 6 mbar, and is mainly composed of carbon dioxide (95%), nitrogen (3%) and trace amounts of oxygen. The water vapour content of the atmosphere is only about 3%. However, the relative humidity of the atmosphere is very high with respect to the low temperature and pressure conditions. This allows indeed the formation of water clouds, but precipitation is not possible. On the surface of Mars no liquid water can be found for the same reason, but nevertheless the existence of old outflow channels and canyons point to the existence of running water in the former geological history of the planet.

With liquid water, a warmer weather and a thicker atmosphere in the past, it is very exciting to think about the possibility of past life existing on Mars. The current environmental conditions do not support life directly on the surface due to the sterilizing effect of the solar UV radiation. However, if life has ever evolved there, it is possible that it still exists in biological niches, sheltered from the radiation beneath rocks or inside the Martian soil, or at the planet’s poles. Large water ice sheets were found at both Martian poles, being partially covered by CO₂ ice during the winter and spring season (Bibring et al. 2004), and also in a crater at high northern latitude (Xie et al. 2006).

During winter time, the northern ice cap extends towards lower latitudes, followed by shrinkage during spring of about 20 km/day, resulting in a minimum extension during the early northern summer. For latitudes higher than 40°, most of the water on the Martian surface is frozen in the polar caps and in permafrost beneath the crust. The permafrost layer gets thicker with increasing latitude, until it rises from the crust at higher latitudes, shaping the ice cap. Topography of the northern and southern polar caps, respectively, was derived from data recorded by the Mars Orbiter Laser Altimeter (MOLA) aboard the Mars Global Surveyor spacecraft (Fishbaugh and Head 2001). The Martian polar caps cover an area of ≈10⁶ km² with a thickness as much as 3–4 km (Clifford 2001). With a supposed age of ≈10⁵–10⁸ years, these layered ice deposits appear to be surprisingly young. However, the Martian ice caps contain therefore a record of the seasonal and climatic cycling of atmospheric CO₂, H₂O, and dust over this period. Therefore, the poles of Mars are within the focus of interest for melting probe applications (Treffer et al. 2006).

Another interesting discovery concerning water ice on Mars was recently obtained from Mars Express data, extending the knowledge of a frozen sea deposit of water ice in the Martian Elysium area (Murray et al. 2006).

3.4 Other places in the solar system

There are, of course, other very interesting planetary bodies in the solar system, where in-situ exploration, supported by melting probe technology shall be considered. There are, however, no mid-term missions defined, where such investigations could be performed.

Europa is not the only Jovian satellite with an icy surface that could be penetrated and
investigated by means of melting. In particular Ganymed is very interesting and many of the arguments presented in Section 3.2 also apply for this moon. Ganymed (as well as Callisto) might also have a subglacial ocean (Spohn and Schubert 2003), but since it would be covered by a much thicker ice layer, the exobiological interest is focussed on its “inner sister”, Europa. 

Titan, Saturns large satellite with a dense atmosphere does also have an icy surface with a high content of hydrocarbons. The spectacular results after the entry and landing of the Huygens probe (Zarnecki et al. 2005; Owen 2005; Lebreton et al. 2005), indicate that penetrating the surface, which may be a crust on liquid ethane/methane by melting could well be appropriate.

Looking even further out in the solar system one may think of applications on the Neptunian satellite Triton, Kuiper belt objects or the many small ice bodies like comets or small icy satellites.

4 Planetary protection aspects

An aspect of space exploration, which has been noticed very early is the effect of human-caused biological cross-contamination between Earth and other solar system bodies (Randolph et al. 1997). A growing awareness about the possible “infection” of other places in the solar system by so called “hitchhiker” bacteria from Earth found its expression in many passionate discussions among the scientific community. These hitchhiker bacteria could colonize on a spacecraft and its equipment leading to a contamination of an extraterrestrial environment. Such a scenario could cause irreversible and dramatic changes of such an alien environment. Furthermore, these alterations could also interfere with scientific exploration and consequently could spoil the measured data. Consequently, the design of all missions to other potentially habitable bodies of the solar system has to consider this aspect of planetary protection.

The measures for planetary protection are defined by international law, based on the outer space treaty of 1967 (Rummel 2001; Rummel et al. 2002).

Techniques, applied for the sterilisation of spacecraft (Engelhardt, 2006, could well be applied for a melting probe, designed to explore the Antarctic lakes. Sterilization is obligatory for any lander mission to Mars or Europa.

5 Conclusions

1. The exploration of Mars’s polar caps and fascinating environments like the putative water ocean below the ice crust of Europa is seen as an important goal for the fields of comparative planetology, astro- and exobiology. Melting probes, already successfully tested in the Antarctic shelf ice, demonstrate that this technology is promising for such applications.

2. Technologies needed for a future Europa mission shall be demonstrated on the ground, leading to an exploration of Antarctic subglacial lakes (like Lake Vostok), which by itself would constitute an important scientific advance.

3. As a first step simple probes as described in this paper are being developed, tested under thermal-vacuum conditions and on Earth (e.g. deep penetration in terrestrial glaciers)

6 Appendix: Thermophysical properties of ices

6.1 Overview
<table>
<thead>
<tr>
<th>Property</th>
<th>Symbol, Unit</th>
<th>H₂O</th>
<th>CO₂</th>
<th>CO</th>
<th>CH₄</th>
</tr>
</thead>
<tbody>
<tr>
<td>Molar mass</td>
<td>M, g/mol</td>
<td>18.01526</td>
<td>44.0098 ± 0.0016</td>
<td>28.01</td>
<td>16.043</td>
</tr>
<tr>
<td>Triple point temperature</td>
<td>Tᵢ, K</td>
<td>273.16 (exact)</td>
<td>216.592 ± 0.001 (ITS-90 secondary ref. point)</td>
<td>68.05 ± 0.05</td>
<td>90.694(1) (ITS-90 secondary ref. point)</td>
</tr>
<tr>
<td>Triple point pressure</td>
<td>Pᵢ, Pa</td>
<td>611.655</td>
<td>(0.51795 ± 0.00010)E6</td>
<td>0.01537(3)</td>
<td>1.169(6) E4</td>
</tr>
<tr>
<td>Critical point temperature</td>
<td>Tₛ, K</td>
<td>647.096</td>
<td>304.128 ± 0.015</td>
<td>132.9</td>
<td>190.6</td>
</tr>
<tr>
<td>Critical point pressure</td>
<td>Pₛ, Pa</td>
<td>2.2064 E6</td>
<td>(7.3773 ± 0.0030) E6</td>
<td>(3.499 ± 0.03) E6</td>
<td>4.592 E6</td>
</tr>
<tr>
<td>Normal melting point</td>
<td>Tₘ, K</td>
<td>273.1525</td>
<td>(Triple point)</td>
<td>68.05</td>
<td>162.0(2)</td>
</tr>
<tr>
<td>Normal boiling point</td>
<td>Tₜ, K</td>
<td>373.124</td>
<td>194.686* (*sublimation pressure = 1 atm, ITS-90 secondary ref. point)</td>
<td>81.60</td>
<td>111.67 (ITS-90 secondary ref. point)</td>
</tr>
<tr>
<td>Density of solid at triple point/melting point</td>
<td>ρₛ, kg/m³</td>
<td>916.700 ± 0.026</td>
<td>1541</td>
<td>919.8</td>
<td>489.8 (400 – 530 over the whole range)</td>
</tr>
<tr>
<td>Enthalpy of sublimation</td>
<td>Hᵥₛ, kJ/kg</td>
<td>2834.359</td>
<td>573.31 at Tₜ</td>
<td>261.3 [CC]</td>
<td>600(23) for 53–90K</td>
</tr>
<tr>
<td>Enthalpy of fusion</td>
<td>Hₘₐₐₙ, kJ/kg</td>
<td>333.44</td>
<td>196.65 at Tₜ</td>
<td>29.86(3)</td>
<td>58.5(2)</td>
</tr>
<tr>
<td>Enthalpy of vaporization</td>
<td>Hᵥₑᵥₘₚ, kJ/kg</td>
<td>2500.5 (0°C)</td>
<td>2255.5 (100°C)</td>
<td>217</td>
<td>510–584</td>
</tr>
<tr>
<td>Specific heat capacity of solid at T₁</td>
<td>cₛ(s), kJ/kg/K</td>
<td>2.2</td>
<td>1.383</td>
<td>1.9</td>
<td>2.73</td>
</tr>
<tr>
<td>Thermal conductivity of solid</td>
<td>λₛ, W/m/K</td>
<td>2.1</td>
<td>0.303 (50 K)</td>
<td>0.4(1) at T₁ (extrapolated)</td>
<td></td>
</tr>
<tr>
<td>Comments</td>
<td></td>
<td></td>
<td>Additional transition at 61.55 K, enthalpy change 22.62(14) kJ/kg</td>
<td>Rotational transition in solid at 20.48 K (ITS-90 secondary ref. point) with λ-peak in cp, enthalpy change 5.81 (1) kJ/kg</td>
<td></td>
</tr>
</tbody>
</table>
6.2 Temperature dependence of selected quantities

6.2.1 Density

With sufficient accuracy from 0 to 273 K:

\[ \rho = 933.31 + 0.037978T - 3.6274 \times 10^{-4}T^2 \text{[kg m}^{-3}] \]

6.2.2 Thermal conductivity

After (Slack 1980), best estimates of \( \lambda \) of Ih \( \text{H}_2\text{O} \) ice between 10 K and the melting point at atmospheric pressure:

\[
\lambda = 619.2/T + 58646/T^3 + 3.237 \cdot 10^{-3}T - 1.382 \cdot 10^{-5}T^2 \text{in W/m/K}
\]

(0.3)

This equation gives about 12% higher values than the (Klinger 1980) equation,

\[
\lambda = \frac{567}{T} \text{W/m/K (T > 50 K)}
\]

6.2.3 Specific heat capacity at constant pressure

\[ C_p = x^3 \left( c_1 + c_2x^2 + c_3x^6 + c_4x^2 + c_5x^4 + c_6x^8 \right) \text{[J/kg/K]} \]

\( x = T/T_f = 273.16 \text{ K and } c_1 = 1.843 \cdot 10^5, c_2 = 1.6357 \cdot 10^8, c_3 = 3.5519 \cdot 10^9, c_4 = 1.667 \cdot 10^2, c_5 = 6.465 \cdot 10^4, c_6 = 1.6935 \cdot 10^6 \) after Haida et al. (1974), Flubacher et al. (1960) and Giauque and Stout (1936).

Note that the pressure dependence of \( C_p \) is only about \( \text{d}C_p/\text{d}P = -8.5e^{-10} \text{T} \) [J/K/kg/Pa] for 0.273.15 K at 1 bar.

6.2.4 Latent heat of sublimation
After Feistel (2006) for 0.273.15 K, approximated by a quadratic polynomial,

\[ L_s = 2636.77 + 1.65924T - 0.00341357^2 [kJ/kg] \]

6.2.5 Viscosity

After Kestin et al. (1978), Hallet (1963) and IAPWS (2003); correlation for temperatures from $-24^\circ C$ [supercooled] to 373 $^\circ C$, at saturation pressure, ± 5%:

\[ \eta(T) = A \left( \frac{T}{B} - 1 \right) ^z \]

\[ A = 1.4147 \times 10^{-4} [Pa \cdot s] \]

\[ B = 226.8 [K] \]

\[ z = -1.5914 \]

7 References (Appendix)


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