CHAPTER 2

Definitions and Background of Periglacial and Permafrost Environments

In literature covering Martian landforms and processes, the terms *permafrost* and *ground-ice* are often used loosely, covering various processes in which ice might have been involved, be it on the surface or in the subsurface. The terms are often used for describing possible sources of water residing as a more or less thick frozen layer in the Martian subsurface or as pore-ice in the Martian regolith and from where it might have been released by processes, such as volcanic heating. Often, landforms are attributed to *permafrost* or are described as *permafrost features* or *ground-ice features* and although the presence of dry Martian permafrost *sensu strictu* cannot be disputed, the search for ground-ice and its distribution in the Martian regolith is still an ongoing process. On Earth, the history of periglacial and permafrost research led to a confusing terminology in the last century, during which primarily former-Soviet, European and US-American scientists worked independently of each other and used different terms to describe essentially similar environmental conditions. Historical aspects of research conducted in periglacial environment and the development of research are extensively covered in, e.g., *Washburn* (1981) or *French* (1996, 2003, 2005) and will not be summarized here. In the following sections, the author focuses on a small selection of relevant work and summarizes some of the aspects that are considered necessary for the understanding of periglacial conditions on Mars. Articles on the general Martian periglacial environments are rare but an almost unmanageable number of contributions deal with the modeling of the stability and distribution of Martian ground-ice, discuss possible Martian ice ages, and describe Martian *permafrost features* at the surface. Here, only literature references concentrating on terrestrial landforms comparable to those on Mars are addressed in upcoming sections. The reader is referred to references cited therein for in-depth discussions.

2.1. Terrestrial Periglacial Environments

According to the nowadays deprecated and obsolete definition by *von Lozinski* (1912), the term *periglacial* refers to the action of frost at margins of glaciated areas during the Pleistocene (*von Lozinski*, 1909, 1912). This definition has been considered to be inadequate as environmental conditions, such as temperature and humidity, are not covered (*Jahn*, 1954). Today, the term is used less restrictedly with respect to location and time. A modern definition of the term *periglacial* contains *"conditions, processes and landforms associated with cold, nonglacial environments [on Earth]"* (*Dylik*, 1964; *Washburn*, 1979a; *French*, 1996; *van Everdingen*, 2005). The term is now widely accepted notwithstanding some criticism regarding its lack of precision (e.g., *Linton*, 1969). Parts of the original meaning of the term *periglacial* as proposed by *von Lozinski* (1912) are nowadays substituted by terms such as *proglacial* (*French*, 1996) or *extraglacial* (*Jahn*, 1975) that generally define areas at margins of ice sheets and glaciers as well as environments under ice-marginal conditions. Early work by Andersson (1906) and Troll (1944) made use of the term *subglacial* as a synonym for periglacial environments although it was argued that the term subglacial refers more to a "vertical climatic-morphologic zonality" and is therefore inappropriate (Jahn, 1975). In Russian literature, the terms *cryology* or *geocryology* are still used frequently today to describe either the periglacial domain (e.g., Yershov, 2004) or processes and environments connected to the action of ice in general. *Cryogenic phenomena and processes* are other terminological alternatives to describe all sorts of ice-related features excluding glacial processes, although these words are not often read today.

In 2005, the Geological Society of London (GSL) published a proceedings volume containing contributions dealing with *cryospheric systems* (Harris and Murton, 2005a). In the volume's preface, cryospheric systems are defined as processes of either periglacial or glacial environments as well as processes at the glacial/periglacial interface. The purpose of that conference was to establish cross-disciplinary work between glaciology and geocryology.

The term *periglacial* refers to immediate results of frost at ground level as well as to cold-climate processes in general, such as aeolian transport and deposition in areas controlled by frost action (Washburn, 1979a). Periglacial landscapes are therefore an expression of climatic conditions and near-surface processes. Scientifically, terrestrial periglacial processes advanced to a widely discussed topic at the beginning of the last century. Geomorphologic features and their significance are summarized in standard textbooks of, e.g., Washburn (1979a), Weise (1983) and French (1996). According to French (1996), there is at least one of two diagnostic criteria which is common to all periglacial environments: The first criterion involves the action of freezing and thawing of the ground with or without contribution of water while the other criterion is the presence of perennially frozen ground or permafrost. French (1996) cites Péwé (1969) who considers all areas underlain by permafrost to be the modern periglacial zone. French (1996) emphasizes that areas of intense frost action are not necessarily underlain by permafrost as observed in various alpine regions which are considered to be periglacial but that do not show any signs of permafrost. Moreover, there are areas of thick permafrost in North America that are not related to the present climatic conditions. However, "the congruence between the extent of the periglacial domain [...] and the global extent of permafrost is remarkably high" as stated by French (1996).

Periglacial landforms and geomorphologic results of periglacial processes cover a wide range of scales ranging from frost-heaving processes on grain-size level resulting in pico-relief landforms to large mass movements and hill-slopes in deglaciated areas at macro-relief scale (for scale classifications see, e.g., Dikau (1989)). Processes in periglacial environments are dominated and characterized by physical weathering with especially frost weathering being the most important process in this context. Physical weathering plays a more important role than chemical weathering in periglacial areas but only the complex of both interacting processes distinguishes cold-climate regions from more temperate ones (French, 1996). As pointed out by Embleton and King (1975) as well as by French (1996), the expressions of periglacial processes are not unique, transitional morphologies exist and the results of these processes can not always be used to infer the correct morphogenesis.

According to Dylik (1964), several processes can still be considered unique for periglacial environments. These processes involve (a) the formation of permafrost, (b) the development of thermal-contraction cracks, (c) thawing and degradation of permafrost, i.e., thermokarst, and (d) formation of wedge- and injection ice. Most processes involve the development of perennially cryotic ground with the formation of segregation ice, frost weathering and disintegration of surface rock by frost wedging, frost heaving, soil creeping, solifluction (more precisely: gelifluction), as well as rock falls and slumps, fluvial processes indicative of seasonally variable discharge, and, finally, aeolian processes (French, 1996).

2.1.1. Controlling Factors of the Periglacial Domain

The three basic, most important and – to a certain degree – independent factors controlling the development of periglacial environments are climate, topography and rock material (*Washburn*, 1979a). These three factors are independent only at a large scale as outlined by *Washburn* (1979a); factors can easily influence each other on a local scale. An additional factor is time which is the most important when paleoclimatic studies are conducted. In contrast to this, dependent factors are considered to be snow cover, vegetation and liquid moisture, which are at most marginally important for the recent Martian environment.

The climate factor has to be taken into consideration at different scales: zonal climate, local climate and microclimate (*Washburn*, 1979a) but only the zonal climate which reflects large-scale effects is considered to be critical to reconstruct past climates and climate change. Climatic parameters controlling periglacial processes are temperature, precipitation, and wind; all of them have to be considered against the background of seasonal changes. The first two parameters are the most essential for climate classifications.

Topography as the second independent factor can influence the local and regional climate but it can also directly influence processes in periglacial environments that are determined by, e.g., slope angles, such as mass-wasting processes (*Washburn*, 1979a).

The third independent factor, rock material, is also influenced by the local climate but it has a stronger influence on the local topography than vice-versa and is therefore considered to be an independent factor. For an in-depth analysis on dependent factors, the author points towards the work by *Washburn* (1979a), p. 13 as this is probably the most complete summary of factors controlling periglacial environments.

2.1.2. Climatic Zonality and Distribution

The definition of the term *periglacial* lacks any generally recognized quantitative parameters. Although some authors define the periglacial domain by the occurrence of permafrost only (*Péwé* (1969) and, with some limitations, also *Tricart* (1967)), many climatic classification schemes have been proposed since the first definition attempts by *Troll* (1944). All of the classifications involve a defined temperature regime as well as a certain mean annual amount of precipitation (e.g., *Peltier*, 1950; *Tricart*, 1969; *Barsch*, 1993). According to *Peltier* (1950), the periglacial morphogenetic region is defined by an average annual temperature range of −15°C to −1°C and an average annual rainfall in the range of 127 mm to 1397 mm; according to *Wilson* (1968, 1969) in *Washburn* (1979a), it is defined by a mean temperature ranging from −12°C to +1°C with annual precipitation in the range of 50 mm to 1250 mm.

Among the plethora of climatic classification attempts for the periglacial zone proposed thus far, those by *Tricart* (1967) and *Tricart* (1969), *Washburn* (1979a); *Barsch* (1993) and *French* (1996) are shortly summarized hereafter.

In contrast to the simple classification attempts by *Peltier* (1950) and *Wilson* (1968, 1969), *Tricart* (1967, 1969) presented a more complex classification scheme which consists of a zone of *cold dry climate with severe winters* which is equivalent to Köppen climate zones D and E, and zones with *cold humid climates with severe winters* that are further separated into arctic (humid parts of the Köppen ET climates) and mountain types. As a third zone, *Tricart* (1967, 1969) suggested a *cold climate zone with small annual temperature ranges* which is found at high-latitude island- and low-latitude mountain areas, see also *Washburn* (1979a) and *French* (1996).

Washburn (1979a) favours the climatic classification as proposed by *Köppen* (1936); *Köppen-Geiger* (1954); *Strahler* (1969) as it provides more precise boundaries and adapts it to the periglacial domain. *Washburn* (1979a) separates four zones consisting of a polar lowlands zone, a subpolar lowlands zone, a middle-latitude lowlands zone and a highlands zone: (1) The *polar lowlands zone* is roughly south of 50°S and north of 55°N with maximum temperatures below 10°C. This zone corresponds to the Köppen ET and EF zones (figure 2.1, table 2.1). (2) The *subpolar lowlands zone* extends roughly from 50°-70°N and 45°-60°S.

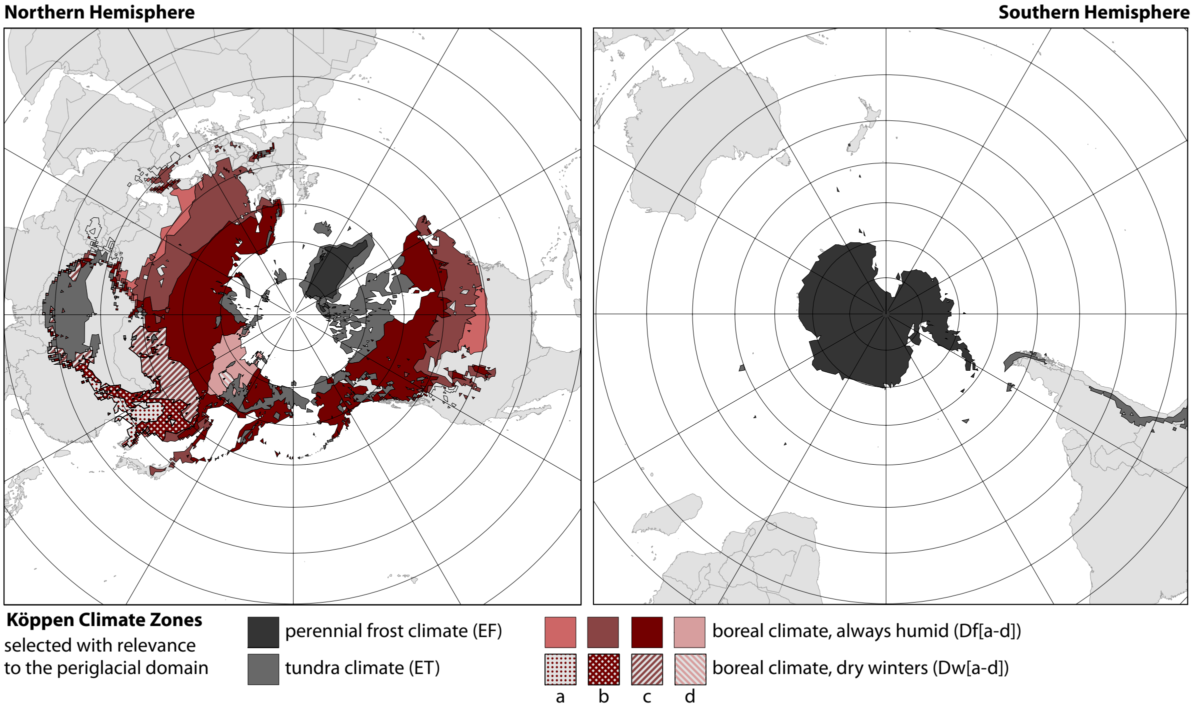

Figure 2.1.: Selected climate zones of the Köppen-Geiger classification (*Strahler and Strahler*, 2005). Selection is based upon the classification of *Washburn* (1979a) for zones characteristic of the periglacial domain. The first letter E represents cold-climate zones with temperatures of the warmest month below 10°C. Climate zones represented by the letter D belong to the boreal climate zone, either humid (f) or winter-dry (w). See text and table 2.1 for additional details. Also see geographic overlap of climatic zones with the permafrost distribution in figure 2.4. Zones have been updated with CRU TS 2.1 temperature and VASClimO v1.1 precipitation data, 1951 to 2000. Grid resolution is 0.5° in lat/lon direction. Polar climate zones extracted from digital data provided by *Kottek et al.* (2006).

Average temperatures range from < −3°C to > 10°C. The subpolar lowlands zone corresponds to the Köppen Dfc, Dfd, Dwc and Dwd climate zones (figure 2.1, table 2.1). (3) The *middle-latitude lowlands zone* with temperatures of the coldest month below −3°C but four months with average temperatures above +10°C extends roughly from 35°-60°N corresponding to the Köppen Dfa, Dfb, Dwa and Dwb climate zones (figure 2.1, table 2.1). (4) The *highlands zone* for which – according to *Washburn* (1979a) – no consistent attempt has been made to adopt a highland zonation, is basically present at all latitudes. For this zone, the critical periglacial boundary is marked by the altitudinal treeline and an altitude of 1000 m has been considered as a minimum value (figure 2.1 and table 2.1).

A more pragmatic type of classification was proposed by *Barsch* (1993) who simply connects periglacial environments with ET and EF climates in the Köppen classification (figure 2.1, table 2.1). *French* (1996) argues that although the classification is consistent with respect to alpine periglacial environments it lacks precision with respect to the lower boundary of arid and semi-arid regions. Barsch's classification also misses the boreal-forest areas that are considered to be important according to *Washburn* (1979a) and *French* (1996).

After proposing a first climatic classification for the periglacial domain which covered four climate zones (*French*, 1976), *French* (1996) modified that approach and also took account of unique plateau locations of the periglacial environment. Except for this addition, this classification scheme is roughly consistent with the one that has been proposed by earlier workers.

Table 2.1.: Selected climate zones after *Köppen* (1936); *Köppen-Geiger* (1954); *Strahler* (1969) with relevance to periglacial environments according to *Washburn* (1979a), see also figure 2.1.

zone	type	subtype	description
D			boreal climate, coldest month <-3°C; warmest month > +10°C
E			cold climate, warmest month <10°C
	f		always humid, zones A-D only
	w		winter-dry, zones A-D only
	T		warmest month between 0°C-10°C, zone E only
	F		all months <0°C, zone E only
		a	hot summers, warmest month >22°C, zones C, D only
		b	warm summers, warmest month <22°C, zones C, D only
		c	short summers, less than 4 months with T >10°C, zones C, D only
		d	extreme cold winters, coldest month with T < −38°C, zone D only

The climatic distribution of the periglacial domain is classified into (1) high-Arctic climates of polar latitudes with strong seasonal and weak diurnal patterns and small annual temperature ranges characteristic of regions such as Svalbard and the Canadian Arctic, (2) continental climates in subarctic latitudes with extreme annual temperature ranges in regions such as Siberia, Alaska, and Yukon, (3) alpine climates of mid-latitudes with well-developed diurnal and seasonal patterns of mountainous areas (Alps), (4) the Qinghai-Xizang plateau (Tibet) to account for this special location at high elevations, and (5) climates of low annual temperature ranges in azonal locations, such as subarctic island climates or mountain climates in low latitudes. These azonal locations are in principle consistent with the third location category proposed by *Tricart* (1967, 1969).

Nonetheless, all of these classifications are considered to be either too restrictive or they do not account for all areas in which, e.g., permafrost occurs (*French*, 1996). Today it is loosely accepted to include all areas where the mean annual air temperature (MAAT) is less than +3° in close connection to the definition given by (*Williams*, 1961). *French* (1996) proposes also to further subdivide the definition by (a) a zone where frost action dominates (MAAT less than −2°) and (b) a zone where frost action occurs but does not dominate (MAAT between −2° to +3°).

This discussion shows that there is much controversy about the definition and zonality of terrestrial periglacial zones; for this reason, a less-quantitative definition of the periglacial domain is favored by many authors. Without exact knowledge about the settings of a particular area it remains problematic to apply any classification scheme.

2.1.3. Periglacial and Glacial Environments

Glaciers belong to the cryospheric domain, a domain of which permafrost and periglacial environments are part and which is defined through all ice-related environments on Earth (e.g., *Harris and Murton*, 2005b). Accordingly, cryospheric processes are connected to the action of ice and cold climates in general, be it as permafrost in periglacial environments or as ice expressed as glacier ice. Any ice-permafrost relationships and interactions are summarized under cryospheric processes which form a comparably small niche in terrestrial research.

Although glacial systems are not a main issue of this thesis, it seems appropriate to shortly emphasize the differences between both systems. Parts of the work described in upcoming sections and also in part III have some relevance to glacial as well as periglacial environments. Nonetheless, complex systems cannot be easily classified into compartments using an off-the-shelf definition and therefore this summary will remain incomplete in many aspects.

In general, glaciers are considered to be extensive bodies of land ice which move downslope under the influence of gravity. The advance and retreat of a glacier is controlled by the rates of accumulation and ablation, and both processes are sensitive to climatic factors such as precipitation and conditions for the recrystallization of ice. An important part of the glacial system is the meltwater environment. Meltwater within or below a glacier affects its behaviour and controls the rate of glacier flow (*Benn and Evans*, 2003). Landscape-forming processes also belong to the glacial system in which erosion of the glacier bed and the proglacial area, transport of sediments and deposition take place. Each of the above mentioned processes and environments are covered extensively in literature (*Benn and Evans*, 2003; *Menzies*, 2002; *Hambrey and Alean*, 2004; *Paterson*, 2001; *Hooke*, 2005)

There are many approaches to classifying different types of glaciers; one classification makes use of sizes, another uses the general shape, again another makes use of the thermal state of a glacier, i.e., *polythermal*, *temperate* or *polar* glaciers, and yet another classifies glaciers according to their confinement. The most important factors for glacier movement however remain the same, be it for small cirque glaciers or for the Antarctic ice cap. It is controlled predominantly by three processes: sliding, deformation of ice, and deformation of the glacier bed (*Benn and Evans*, 2003).

While the glacial system is characterized and defined by the body of ice itself as well as all glacier-related processes connected to it, the periglacial domain is defined by the climatic environment. It is imaginable (and often observed) that certain landforms can coexist and even develop under similar conditions in both domains. However, although the definitions are straightforward, there are landforms for which a clear classification remains problematic, such as the separation of cold-based glaciers from the glacial domain. According to the definition of the periglacial environment they belong to the periglacial domain although all major glacial textbooks treat such features within a glacial context. There has been a significant amount of research performed by authors on the occurrences of cold-based glacial systems near the Martian equator (*Head and Marchant*, 2003; *Shean et al.*, 2005a; *Milkovich et al.*, 2006). Some contributions are briefly summarized in section 3.2.3.

The semantic discussion regarding glacial and periglacial systems culminates in the discussion of the nature of rock glaciers. The term *glacier* implies a glacial origin, and glacier-like deformation and movement of a rock glacier body. Two factors are treated as critical when the nature of rock glaciers is discussed. First, the thermal regime as outlined below and secondly, the "influence of glacier ice motion on the dynamics of the rock glacier" (*Harris and Murton*, 2005b). Glaciers which are partly debris covered are no rock glacier by definition as they contain slowly melting glacier ice which cannot be part of the permafrost environment. On the other hand, a cold-based glacier moves by internal friction only and can also not be considered as a rock-glacier system whose movement is controlled by gravitational creep (*Harris and Murton*, 2005b). Nevertheless, there are environments where rock glaciers exist below, e.g., cirque glaciers, therefore the boundary between both systems is diffuse. More detailed remarks on rock glaciers and their nature are provided in section 3.

The essential concept for separating periglacial and glacial landscapes is the thermal regime. As summarized by *Jahn* (1975), glacial systems occur in humid areas with low temperatures. In contrast to this, periglacial environments are found in arid climatic zones. Periglacial climates differ significantly from glacial climates and it is a mistake to treat the periglacial climate as a predecessor to more severe glacial climates (*Jahn*, 1975).

2.1.4. Terrestrial Permafrost Environments

Permafrost is the most diagnostic feature of periglacial environments although its presence is not required for an environment to be called periglacial (e.g., *French*, 1996). In principle, permafrost is a perennially frozen layer of variable thickness that has formed as a consequence of long periods of winter-cold and short periods of thawing during summer

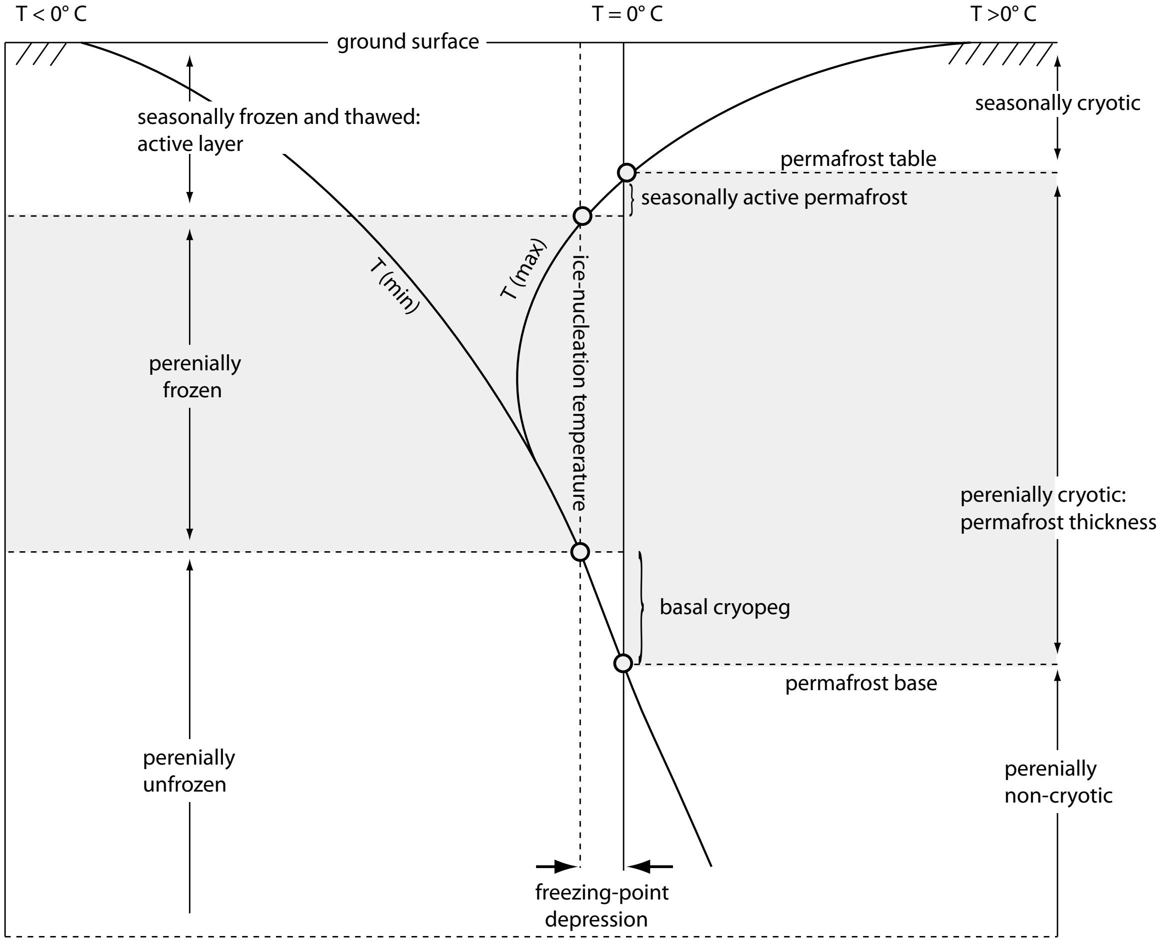

Figure 2.2.: Permafrost terminology and definitions of perennially (un-)frozen or thawed ground with respect to temperature regime, modified after (*van Everdingen*, 1985; *Associate Committee on Geotechnical Research (ACGR)*, 1988).

(*French*, 1996). Terminologically as well as genetically, the permafrost zone is therefore a subset of the periglacial zone. The terms *permafrost*, so-called *perennially frozen ground* or *pergelisol* are treated slightly differently in literature and different definitions are connected to each of the terms. The first English-language definition describes permafrost as permanently frozen ground in areas of cold climates (*Muller*, 1947).

A more common and precise definition describes permafrost as consolidated or unconsolidated ground which is characterized by a temperature of lower or equal to 0°C over at least two subsequent years (*van Everdingen*, 2005). According to this definition, ground ice or perenially frozen ground is not required for an environment to be called permafrost because included pore water could still be liquid at temperature ranges below 0°C as the actual physico-chemical state of water is controlled by the content of dissolved salts. To account for this, the temperature-related definition has been expanded by a description of state which introduces the terms *cryotic* and *non-cryotic* (figure 2.2). These terms allow a description based upon the frozen or unfrozen state of material, independent of the contents of water or ice.

Perennially cryotic ground is therefore a synonym for permafrost in which permafrost is either unfrozen, partially frozen or completely frozen. Consequently, according to that definition, perennially frozen ground is a synonym for permafrost but per-

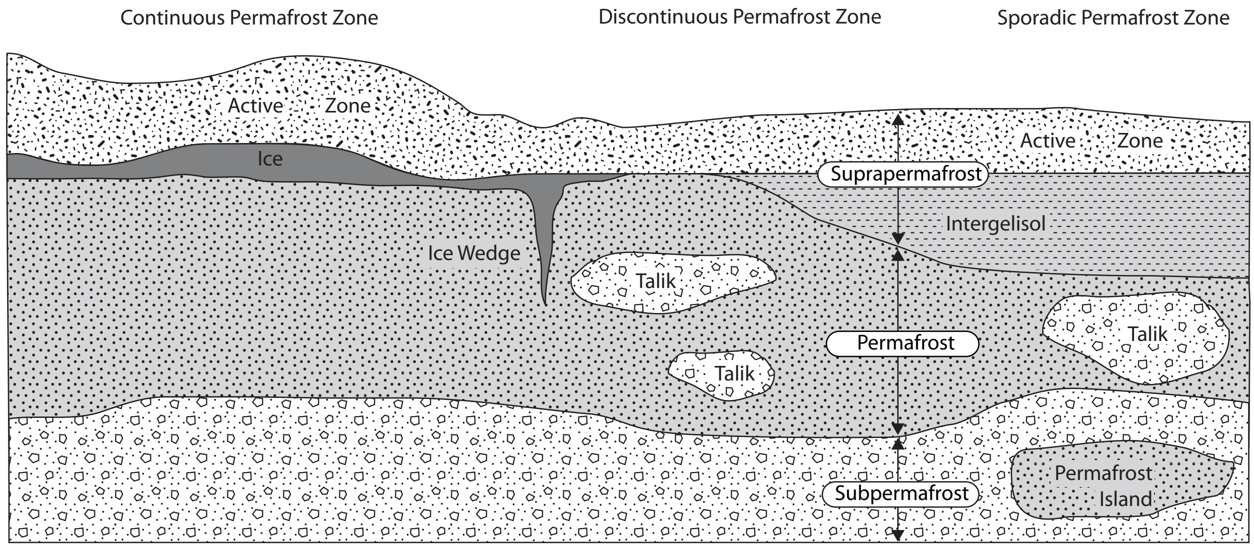

Figure 2.3.: Nomenclature of structural features in terrestrial permafrost. The active layer extends to the continuous permafrost zone (left). Towards the right half of the figure, permafrost becomes discontinuous with isolated patches of unfrozen ground and taliks. Modified after *Stearns* (1966) in *Washburn* (1979a), p. 29.

mafrost is not necessarily perennially frozen. Water is – independent of its physical state – no prerequisite to permafrost. In general, if ice is present, the wording *ice-rich permafrost* is used in order to separate it from *ice-free* or *dry* permafrost.

The thickness of permafrost depends on the thermal conductivity (~1/geothermal gradient) of the subsurface material and the near-ground surface temperature; more precisely, it is controlled by the internal heat gain with depth on one hand and heat loss from the surface on the other. Furthermore, it depends on the material density, pore volume, and grain sizes. The thickness of permafrost can vary a few centimeters from year to year and reach thicknesses of up to 1500 meters (e.g., Resolute Bay, NWT, Canada with 390 m; Tiksi, FSU with 630 m; Schalagonzy, FSU with 1500 m, values according to *Embleton and King* (1975), p. 30 and references cited therein). The influence on the geothermal heat flux on, e.g., the degradation of permafrost has been discussed by *Washburn* (1979a) on the basis of *Terzaghi* (1952). *Embleton and King* (1975) summarize that permafrost temperatures are lowest at the surface and are influenced by temperature cycles at the surface. The depth that is affected by surface temperature variations depends on the timescales: while diurnal changes in temperature affect the upper layer to a depth of max. 1 m, seasonal changes can affect permafrost to a depth of up to 15 m. However, it is not possible to establish a global connection between the distribution of permafrost on the basis of surface or mean air temperatures alone as too many factors influence the distribution of permafrost.

Most of the geomorphologic activity in permafrost is restricted to the *active layer*, i. e., the seasonally thawed upper layer of the subsurface. Therefore, this layer is the area in which processes such as chemical transport, flow of groundwater and mass movements take place and which plays the most relevant role in shaping of the periglacial landscape. Seasonal thawing and freezing cycles of the active layer have several consequences: freezing of water in the active layer at the beginning of winter seasons results in formation of lenticular and segregation ice with the consequence of frost-heaving processes and the associated destabilization of hillslopes (e.g., *Embleton and King*, 1975; *French*, 1996). Upon thawing, destabilized slopes result in gravitationally driven mass transport. Beyond

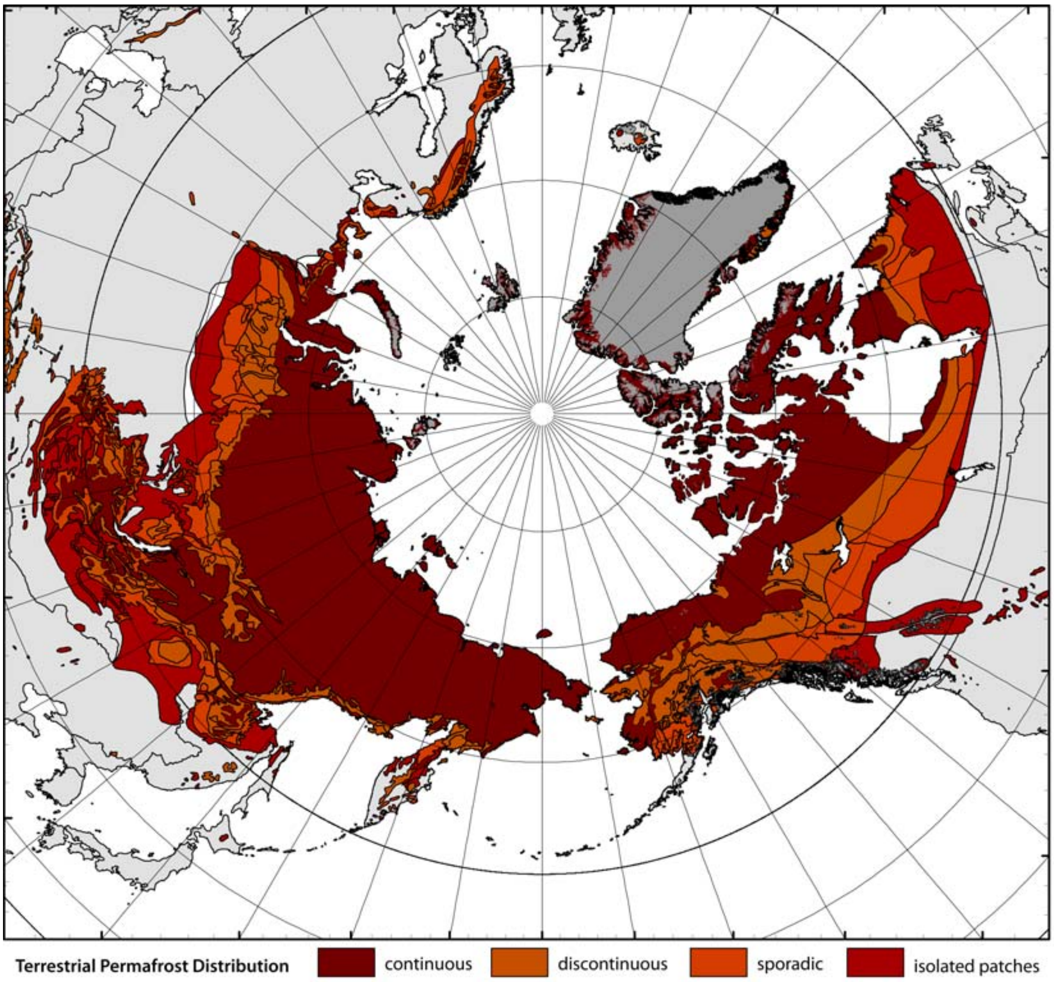

Figure 2.4.: Distribution of continuous (90%-100%), discontinuous (50%-90%), sporadic (10%-50%) and isolated (0%-10%) permafrost on Earth, spatial distribution 180°W-180°E, 25°N-90°N. Sub-classes of overburden thicknesses have been combined appropriately by the author for simplicity reasons. Modified after *Brown et al.* (1997).

that, compaction caused by surface layers increases pore-water pressures and is responsible for the destabilization of the soil promoting mass movement processes.

Hydrologic processes and groundwater transport in areas of permafrost differ significantly from processes in areas without permafrost. Groundwater discharge and exchange of material with the surface layer are limited to taliks, i.e., isolated unfrozen bodies in permafrost environments (*French*, 1996) (figure 2.3). The volume of pore water as a main component of ice-rich permafrost is far higher under frozen conditions than under thawed condition, so that subsidence of thawed ground and formation of thermokarst landforms are favored as soon as the ground becomes unfrozen.

The thickness of the active layer depends on a variety of factors, such as mean air temperature as well as temperature fluctuations, periods of temperature variations, thermal diffusivity of soil (type of soil), vegetation, drainage, water content, snow cover, slope angle and slope exposition. A rough approximation of the thickness of an active layer has been provided by *Gold and Lachenbruch* (1973) (as cited in *French* (1996)). Its basis is defined by the mean annual temperature which must be ≤0°. However, in cases of relict permafrost, it is common to observe an active

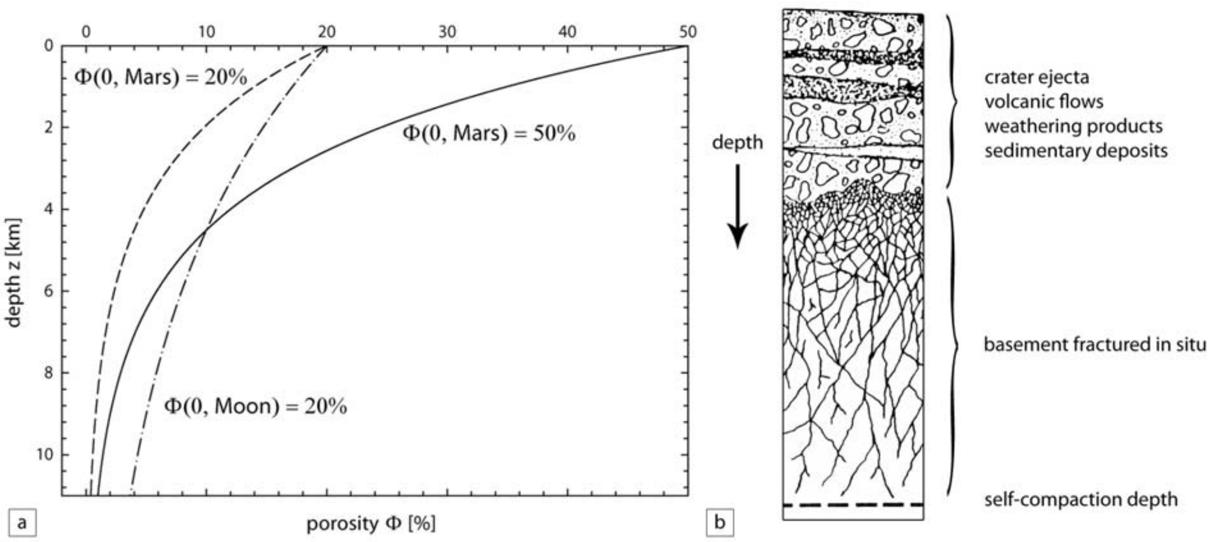

Figure 2.5.: Idealized vertical profiles of the Martian crust, [a] porosity Φ vs. depth z according to equation 2.1 for the Martian case with a surface porosity of Φ_0 =20% and Φ_0 =50% and a decay constant K =2.8 km (*Clifford*, 1981) in comparison to the Lunar megaregolith with Φ_0 = 20% and a decay constant K = 6.5 km (*Binder and Lange*, 1980), after *Clifford* (1981) and *Clifford* (1984) in *Squyres et al.* (1993); [b] idealized stratigraphic column of the Martian crust, after *Clifford* (1981).

layer which is separated from the main permafrost body by a layer with temperatures above 0° (e.g., *Embleton and King*, 1975; *French*, 1996).

Permafrost on Earth occurs in two overlapping regions. One region is connected to high latitudes (*polar permafrost*), the other is connected to locations of higher altitudes, be it alpine environments (*alpine permafrost*) or plateaus (*plateau permafrost*) (figure 2.4). Additionally, extensive terrestrial permafrost areas are situated below sea level on the continental shelves of the Arctic sea. All together, 24% of the land subsurface is considered to be underlain by permafrost (*Baranov*, 1959; *Shi and Mi*, 1988), of which 50% are in the FSU (*Black*, 1954).

In general, permafrost is classified according to its distribution in the subsurface as follows: A *continuous permafrost* distribution is defined as permafrost that occurs at all localities, except for very localized zones of unfrozen ground below, e.g., lakes, such as taliks. In contrast to this, *discontinuous permafrost* is defined by permafrost bodies which are separated by areas of unfrozen ground. The end member in this sequence is defined by so-called *isolated permafrost* where isolated islands of permafrost occur in a generally unfrozen body (*Embleton and King*, 1975) (figure 2.4).

Details on the distribution and various appearances of ground-ice are extensively covered in, e.g., *Embleton and King* (1975); *Washburn* (1979a, 1985); *Church and Slaymaker* (1985); *Barsch* (1993); *French* (1996) and *Yershov* (2004); the connection between permafrost features and climate change is discussed in *Washburn* (1979b).

2.2. Martian Permafrost and Ground Ice

The climatic classifications summarized in section 2.1.2 show that there is still much controversy about the climatic zonation and distribution of periglacial environments on Earth. Notwithstanding any doubts connected to one or the other classification, it becomes obvious that Mars can be considered as a planet

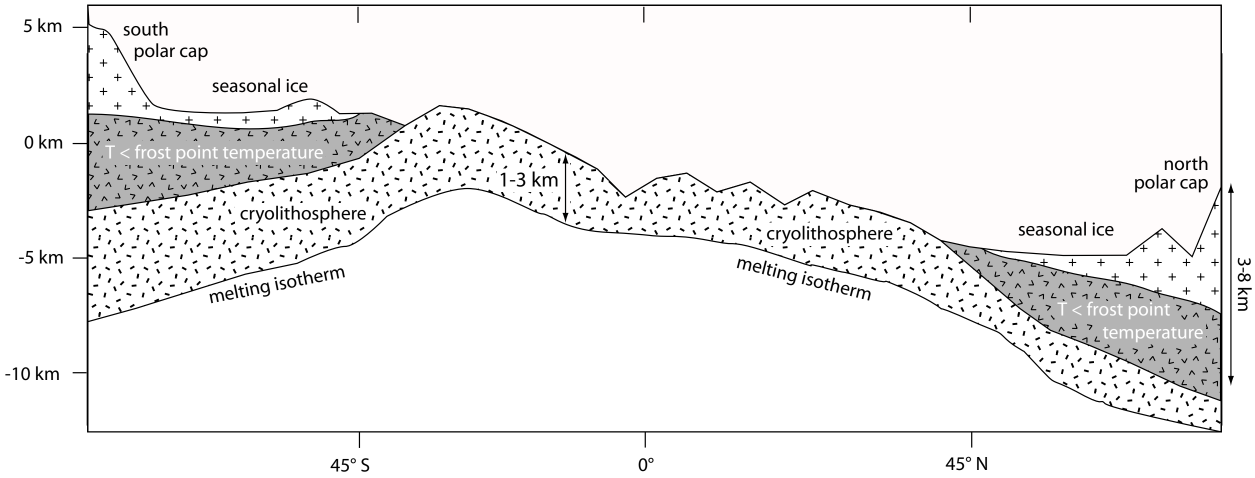

Figure 2.6.: Theoretic cross-section profile of the Martian crust. Profile along 0° longitude from south to north pole showing a 5°-interval MOLA topographic profile of the upper Martian crust and the latitudinal variation of the 273 K isotherm. Ground ice is stable and in equilibrium with the atmosphere where temperatures are below the frost point of atmospheric water vapor (*Farmer and Doms*, 1979) or where it is isolated from the atmosphere by a regolith cover (*Clifford and Hillel*, 1983; *Fanale et al.*, 1986); figure on the basis of *Squyres et al.* (1993) after *Fanale* (1976); *Rossbacher and Judson* (1981).

which is controlled by the periglacial domain at all locations on the surface at all times.

Strong seasonal and diurnal temperature variations are comparable to the harsh and hyperarid ET and EF climates (figure 2.1, table 2.1). The lack of precipitation in recent time has not promoted development of surficial periglacial processes that are common on Earth and that require at least a certain amount of rain- or snowfall. The applied classifications allow the location of environments on the Earth which could show morphologies that are comparable to the Martian environment, such as the Dry Valleys in Antarctica or the high Canadian Arctic.

Apart from the wealth of surface morphologies on Mars indicative of periglacial and permafrost environments, modelling work has significantly contributed to the understanding of the distribution of Martian ground ice and its stability in connection to orbital oscillations (e.g., *Mellon and Jakosky*, 1995). The question of ground-ice stability must be seen under (a) present and (b) past conditions, where a higher obliquity of the planet's spin axis (e.g., *Murray et al.*, 1973; *Pollack*, 1979; *Toon et al.*, 1980; *Jakosky et al.*, 1995) might have caused re-deposition of polar volatiles in equatorial latitudes (*Levrard et al.*, 2004) and affected the stability and distribution of ground ice (e.g., *Mellon and Jakosky*, 1995). Modelling of the tilt of the spin axis suggests that the average axis tilt over the last 5 Ma was almost 40° (*Laskar et al.*, 2004).

For Mars, the term *cryolithosphere* is often used as synonym for the term *permafrost, sensu lato*, similar to its usage in Russian literature for terrestrial systems (*Kuzmin* (1977) as cited in *Squyres et al.* (1993), *Kuzmin* (2005)); it includes also the surficial Martian polar caps. The cryolithosphere is part of the cryosphere which extends from the subsurface to the atmosphere at heights of up to 140 km (*Fanale*, 1976; *Kuzmin*, 2005). *Kuzmin* (2005) emphasizes that, apart from water ice, carbon dioxide (CO_2), the main atmospheric constituent, might be present either as solid CO_2-ice, as liquid or as gaseous hydrate ($CO_2 \cdot 6H_2O$). An extensive discussion about the role of CO_2 in the cryolithosphere is provided by, e.g., *Kuzmin* (2005). The following section concentrates on the core physical properties of the Martian regolith and its capacity to store water-ice.

For the present conditions on Mars, ground ice is not only controlled by the atmosphere but also by the composition of the so-called Martian megaregolith and the exchange of volatiles between both systems. Similarly to the Moon (*Hartmann*, 1973, 1980), the Martian crust is heavily modified by impact processes and ejected debris material of considerable thickness interbedded by lava flows (*Fanale*, 1976; *Carr*, 1979) (figure 2.5a-b). This porous medium, the megaregolith, is known from the Moon and appropriate scaling to Martian gravitational conditions (*Clifford*, 1981) resulted in estimates of the so-called decay constant with K = 2.8 km which controls the porosity Φ at depth z (figure 2.5a) according to

$$\Phi(z) = \Phi(0)^{(-z/K_{\text{Mars}})} \tag{2.1}$$

The self-compaction depth is the depth at which the porosity becomes smaller than 1% (figure 2.5a-b) and is located at approximately 8.5-11 km, depending on the value for the initial porosity (usually 20% to 50%). This estimate amounts to a pore volume of 8-20×10^7 km^3 in which water/ice can be stored and which amounts to an equivalent of a global ocean with a depth of 550 m to 1400 m (*Clifford* (1981) and *Clifford* (1984) as cited in *Squyres et al.* (1993)). The reader is referred to the discussion on factors influencing the porosity in *Squyres et al.* (1993), p. 526.

It is considered that temperatures are below freezing immediately below the surface although surface temperatures can reach values that exceed 273 K on Mars. The depth z of the cryolithosphere's lower bound is calculated by

$$z = k \cdot \frac{T_{\text{mp}} - \bar{T}_{\text{s}}}{Q_{\text{g}}} \tag{2.2}$$

with k as the thermal conductivity of the regolith, T_{mp} is the melting-point temperature, $\bar{T}_{\text{s}}$ is the mean annual surface temperature and Q_{g} is the geothermal heat flux (*Fanale*, 1976). *Squyres et al.* (1993), p. 527 states that the mean annual surface temperatures are known to be in the range of ~160-220 K ± 5 K, but all other variables have uncertainties of 25-100%. The thermal conductivity of terrestrial permafrost has been briefly qualitatively discussed in section 2.1.4 and the same constraints on the thermal conductivity also apply for the Martian environment. For a discussion on error ranges and the derivation of sensible values, see *Squyres et al.* (1993), p. 528 and *Clifford and Fanale* (1985). Values for the calculation of the depth of the Martian cryolithosphere according to equation 2.2 are k = 2Wm^{-1}K^{-1} (*Clifford and Fanale*, 1985), T_{mp} is considered to be between 252-273K, depending on the amount of dissolved salt (*Squyres et al.* (1993) and references cited therein), and Q_{g} is ~3×10^{-2}Wm^{-2} (*Fanale*, 1976). According to equation 2.2, the base of the cryolithosphere is estimated to be in the range of 1-3 km (1-2 km (*Kuzmin*, 2005)) at the equator and 3-8 km (5-6 km (*Kuzmin*, 2005)) at the poles (figure 2.6). The depth increases towards the poles due to a latitudinal decline of mean annual surface temperatures.

Ground ice is stable in the subsurface at locations where it is in equilibrium with the atmospheric water vapor content. Although mean annual surface temperatures of 216 K (e.g., *Kuzmin*, 2005) are below the freezing temperature of water, the frost-point temperature corresponding to an annual average water vapor amount in the atmosphere of ~12μm is about 198 K (*Farmer and Doms*, 1979) or as high as 203 K (*Paige*, 1992) according to the relation

$$M(T) = \rho_0(T) \int_0^h e^{-zH} dz, \tag{2.3}$$

M is the total normalized mass of water in the vertical column, ρ_0 is the density of water vapor at the surface, h is the height to which water is mixed, z is the height and H is the scale height of water. This results in a frost-point temperature of 198 K with a well-mixed atmosphere of 10 pr μm (figure 2.7). The occurrence of (sub)surface ice is therefore restricted to areas below that frost point and this results in a distribution polewards of ±40°N/S (e.g., *Farmer and Doms*, 1979; *Fanale*, 1976) or ±45°N/S according to *Kuzmin* (2005). The frost-point temperature estimates rely additionally on surface albedo and surface thermal inertia (e.g., *Paige*, 1992; *Mellon and Jakosky*, 1995) which is defined as

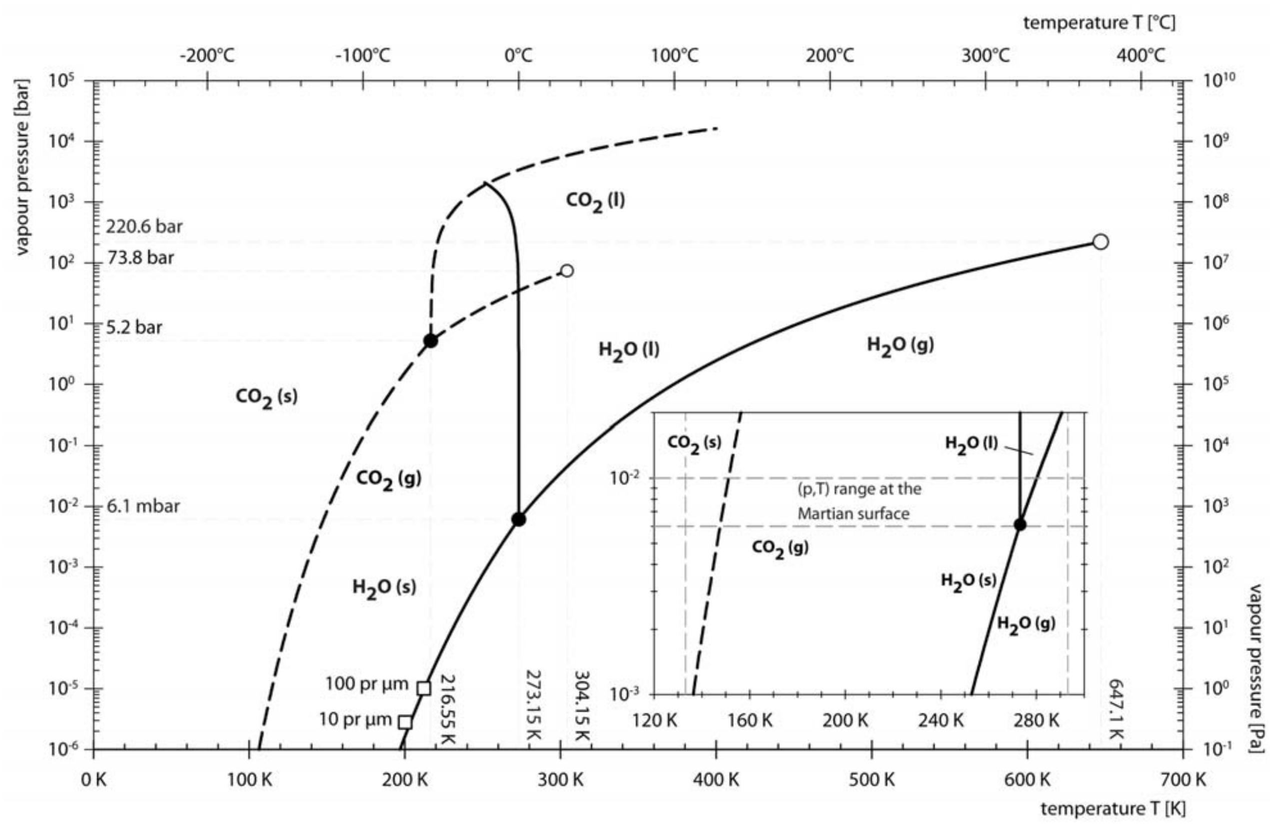

Figure 2.7.: Phase diagram for water and carbon dioxide as well as frost-point temperatures for a well-mixed atmosphere and 10 pr μm and 100 pr μm atmospheric water vapor; filled and empty circles are critical and triple points, respectively; see text for explanations; data source *Wagner et al.* (1994, 2000); *ChemicaLogic* (1999).

$$I = \sqrt{k\rho C} \quad [\mathrm{J \cdot m^{-2} s^{-1/2} K^{-1}}] \qquad (2.4)$$

k is the thermal conductivity measured in $\mathrm{Wm^{-1}K^{-1}}$, and C is the specific heat capacity measured in $\mathrm{Jkg^{-1}K^{-1}}$.
Due to large variations of surface thermal inertia values on Mars ranging from 24-800 $\mathrm{Jm^{-2}s^{-1/2}K^{-1}}$ (*Jakosky et al.*, 2000) which are additionally controlled by existing ground ice (*Paige*, 1992) as well as albedo variations, local differences in ground-ice stabilities seem to be not an exception but the rule. For in-depth discussion on this topic the reader is referred to *Fanale* (1976); *Farmer and Doms* (1979) and *Paige* (1992).
Clifford and Hillel (1983) and *Fanale et al.* (1986) performed analyses on the long-term stability of ground ice in the regolith and estimated that at low latitudes, half of the ground-ice covered by a 100 m dry layer of soil would have been lost during the last 3.8 Ga (*Clifford and Hillel*, 1983). More detailed work showed that at low latitudes ground ice would have been completely sublimed to a depth of a few hundred meters and to a few tens of meters at higher latitudes over the last 3.8 Ga (*Fanale et al.*, 1986).

The plethora of ice-related landforms described from equatorial latitudes must therefore be relics and witnesses of a past climate. Ice must have been preserved due to "diffusion-limiting properties of a fine-grained regolith" and a desiccated upper layer with a thickness of several 100 m (*Fanale et al.*, 1986; *Squyres et al.*, 1993); see, e.g., section 3.2.3 and chapter 12. This conclusion again makes scenarios of changes in the tilt of the Martian rotational axis plausible (*Murray et al.*, 1973; *Pollack*, 1979; *Toon et al.*, 1980; *Jakosky et al.*, 1995; *Mellon and Jakosky*, 1995).

2.3. Selection of Landforms

Periglacial environments host a variety of landforms ranging from millimeter-scale to several hundreds of meters and more. Consequently only few landforms can be investigated on the basis of space-borne remote sensing data. Although small-scale processes are usually responsible for macro-scale landforms, e.g., formation of needle-ice and creep processes at grain-level often result in large-scale gravitational slope movements and gelifluction processes, the focus here is put on periglacial landforms that have a size to be detected from Mars-orbiting spacecraft and that have been discussed in research history. These landforms are related to (a) the creep of mountain permafrost, i.e., gelifluction and rock-glacier analogues, (b) formation and distribution of thermal contraction polygons, and (c) formation of frost mounds. The focus of these landforms is put on investigations of rock-glacier analogues as the amount of literature since the late 1970s is far more extensive when compared to work on thermal contraction polygons or frost mounds that usually require high-resolution image data which did not become available until the end of the 1990s.

Most terrestrial investigations of such landforms are performed in tedious field work and analysis of temporal data that aid understanding of the formation conditions and distribution. This kind of field data is not available from Mars and space-borne instruments do usually not reach the required resolution.

The summary on terrestrial landforms discussed hereafter can only provide a short overview of modern work but will focus not only on information obtained from aerial investigation but also on those pieces of information that have been gathered through field work and that seem to be relevant for the understanding of Martian landforms. □