Chapter 2
Antarctic Permafrost Soils

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2.1 Introduction

Antarctica, with an area of 14 million km^2, is the worlds largest continent, yet exposed ground on which permafrost soils occur covers a mere 49,000 km^2, or about 0.35% of the entire continent (Fox and Cooper 1994). The continent is roughly circular in outline, and its topography is dominated by two massive ice sheets (Fig. 2.1); the East Antarctic Ice Sheet with an average elevation of around 3,000 m, and the West Antarctic Ice Sheet with an average elevation of around 1,500 m. A major physiographic feature is the Transantarctic Mountains, which extend over 3,500 km and separate the two ice sheets. Bare ground areas are found scattered around the margin of the continent where the ice sheets have thinned or receded, in the Antarctic Peninsula and along the Transantarctic Mountains (Fig. 2.1). The largest ice-free area is in the Transantarctic Mountains (23,000 km^2 estimate), which includes approximately 7,000 km^2 in the Dry Valley region, the largest contiguous area of bare ground.

The climate for formation of soils and permafrost throughout Antarctica is severe. With very low mean annual temperatures, negligible effective precipitation and rare occurrences of mosses and lichens, except for the Antarctic Peninsula where plant life including some grasses are more abundant, the soils have aptly been described as Cold Desert Soils (Tedrow and Ugolini 1966; Campbell and Claridge 1969). The exposed landscapes are dominated by glacial valleys with land surfaces and deposits that show the influence of glacial activity, which has extended from the Late Pleistocene to earlier than Miocene times (Denton et al. 1993; Marchant et al. 1993). Notwithstanding the tiny proportion of the continent that is ice-free and exposed to weathering processes, a large degree of diversity is found in both the soils and permafrost, owing to the wide variations in the environmental and geomorphic forces.

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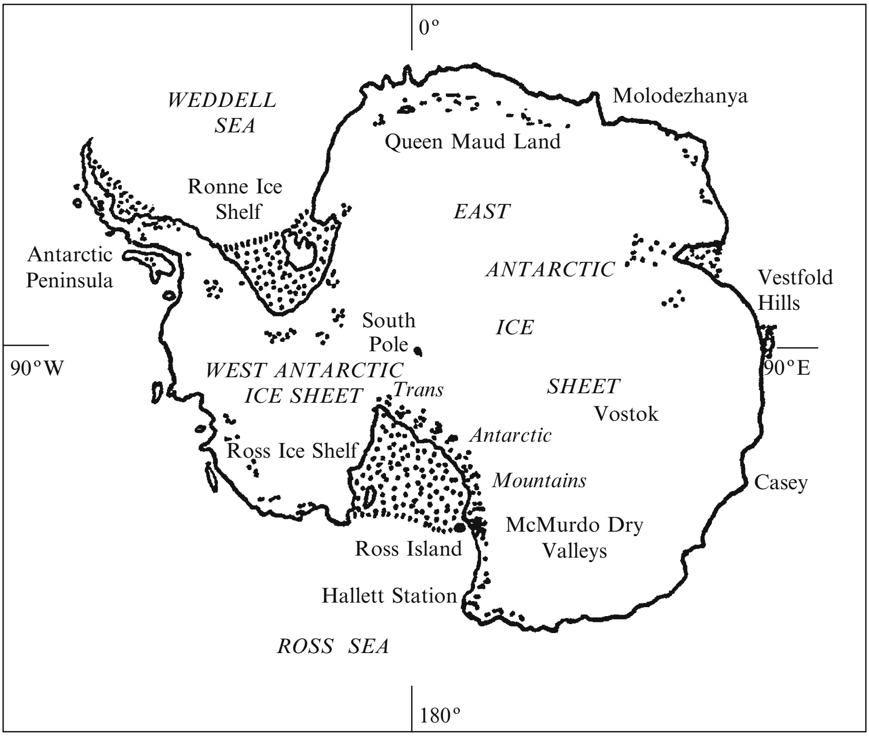

Fig. 2.1 Location map with areas of ice-free ground (not exact or to scale)

2.2 The Climatic Environment

The climate of Antarctica embraces the most extreme cold conditions found on Earth. Antarctica is cold because the solar radiation is only 16% of that at equatorial regions, and also because of the high average surface elevation of the ice sheet, which in places exceeds 4,000 m. Temperatures as low as −89°C have been recorded at Vostok (Fig. 2.1), and −49°C at the South Pole. However, mean annual air temperatures increase nearer the coast where land is exposed, and in the northernmost areas (−25°C at Mt. Fleming at the head of Wright Dry Valley near the edge of the Polar Plateau, −20°C at Vanda Station in the Dry Valleys, −18°C at McMurdo Station on Ross Island, −15°C at Hallett Station). Further north, in coastal areas of East Antarctica, warmer climates are found (MacNamara 1973; Burton and Campbell 1980). At Davis Station in the Vestfold Hills, mean annual temperature is −10.2°C, while at Molodezhnaya and Casey (Fig. 2.1) similar temperatures to those at Davis Station are experienced.

Air temperatures directly influence permafrost properties, with the active layer thickness decreasing from around 80 to 100 cm in the warmer coastal and northern regions to 2 cm or less in the cold inland high-elevation sites (Fig. 2.2) following the

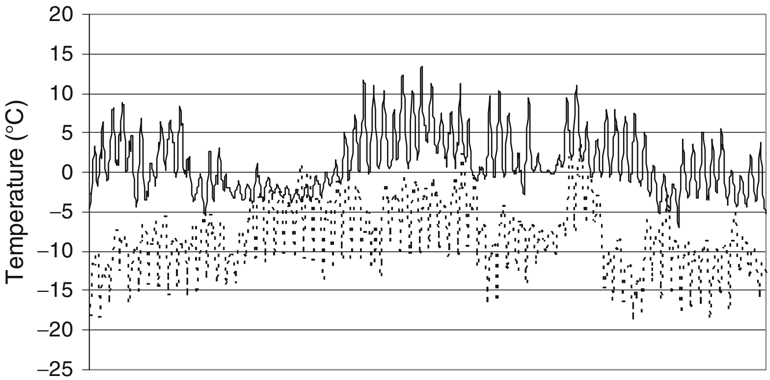

Fig. 2.2 Hourly temperature records from Marble Point (*solid line*; 70 m above sea level (asl), measurement at 7.5 cm) and Mount Fleming (*dashed line*; 2,000 m asl, measurement at 2 cm) from December 4 2002 to February 12 2003. The records illustrate the large difference that site climate has on soil thermal properties

adiabatic lapse rate (Campbell and Claridge 2006). Other soil thermal properties related to geographic differences in climate include the length of the thaw period, the number of thaw days during summer, the number of freeze/thaw cycles that occur and the length of time that the soil may be continuously above freezing. At Marble Point, for example [approximately 70 m above sea level (asl) and permafrost table at 60 cm], the thaw period (measured at 7.5 cm depth) extended over 70 days, there were 34 freeze–thaw cycles and 16 days when the soil temperature was continuously above 0°C (Fig. 2.2). By contrast, at Mt. Fleming (2,000 m asl, permafrost table approximately 2 cm) the thaw period, measured at 2 cm depth, extended over 31 days, but with only 6 days in which soil temperature was briefly above 0°C.

The mean annual precipitation over Antarctica averages around 50 mm per year, with least falling inland and most in coastal locations. In the McMurdo Dry Valleys, one of the driest areas of Antarctica, precipitation averaged 13 mm per year on the valley floor near Lake Vanda and 100 mm per year in nearby upland mountains. Around the periphery of East Antarctica, precipitation is much higher, with 650 mm per year at Molodezhnaya in Enderby Land (MacNamara 1973). The precipitation normally falls as snow, and little is available for direct soil moistening because of ablation and evaporation. Despite the minimal amounts of soil moistening, distinct soil climate zones, based on moisture availability, have been recognized (ultraxerous, xerous, xerous to subxerous, oceanic subxerous and moist zones; Campbell and Claridge 1969). Soils of the ultraxerous zone are found in arid inland areas, rarely if ever have liquid water present, and have ground temperatures that are seldom above freezing point. At the other extreme, moist soils in coastal environments may be moistened at the soil surface, and ground temperatures remain above freezing point for periods throughout the year.

In Antarctica, the soil climate and permafrost properties are strongly influenced by the surface radiation balance, since the soil thermal regime is consequent upon the gains and losses of radiation from the soil surface. Surface radiation balance investigations for soils at several sites were reported by Balks et al. (1995), MacCulloch (1996) and Campbell et al. (1997), who found that soils with dark-coloured surfaces had low albedo values (approximately 5% at Scott Base) while soils with light-coloured surfaces had much higher albedo values (26% at Northwind Valley). Differences such as these, when coupled with available soil moisture, translate into appreciable differences in the diurnal soil thermal regime and permafrost characteristics. At Bull Pass in Wright Valley, for example, a soil surface with approximately 50% dark-coloured clasts had summer soil temperatures (measured at 2 cm) up to 5°C higher (max 17°C) than in adjacent soil with a light-coloured surface, while the mean annual soil temperature at that depth was 0.25°C greater than for the light coloured soil.

2.3 The Geologic Environment

The Antarctic plate, like other parts of Gondwanaland, is formed mainly from Precambrian to Lower Paleozoic basement rocks, intruded by granites and peneplained by weathering and glacial erosion with overlying sediments of sandstones, siltstones, coal measures and tillites. Jurassic basic igneous rocks were intruded to form widespread sills. Other more recent volcanics occur along major orogenic zones. The present glacial environment is believed to have established after the separation of Antarctica from South America which allowed the formation of a circumpolar circulation pattern.

Antarctic soils and permafrost occur in a geological setting where the time scale for landform development and weathering processes extends back to the Miocene or earlier, and in which the glacial events responsible for till deposition are related to several distinct sources. They include glaciations related to the East Antarctic Ice Sheet, the West Antarctic Ice Sheet and to Alpine glaciers (Figs. 2.3a and 2.3b).

2.3.1 The Glaciological Setting

The East Antarctic Ice Sheet is believed to have been stable since Miocene times (Denton et al. 1993; Marchant et al. 1993; Sugden et al. 1993). Evidence from dated $^{40}Ar/^{39}Ar$ in situ volcanic ashes occurring in association with soils from unconsolidated tills in the Dry Valleys, from basaltic flows interbedded with widespread tills and from reworked clasts in moraine sequences, indicate that there has been no significant expansion of this ice sheet or landscape evolution at least since mid-Miocene times. The West Antarctic Ice Sheet has a different history. It rests on bedrock mostly below sea level, and is dramatically affected by sea-level changes. There is clear evidence that during low sea levels, the associated ice shelves grounded and expanded, causing ice to flow backwards into valleys along the

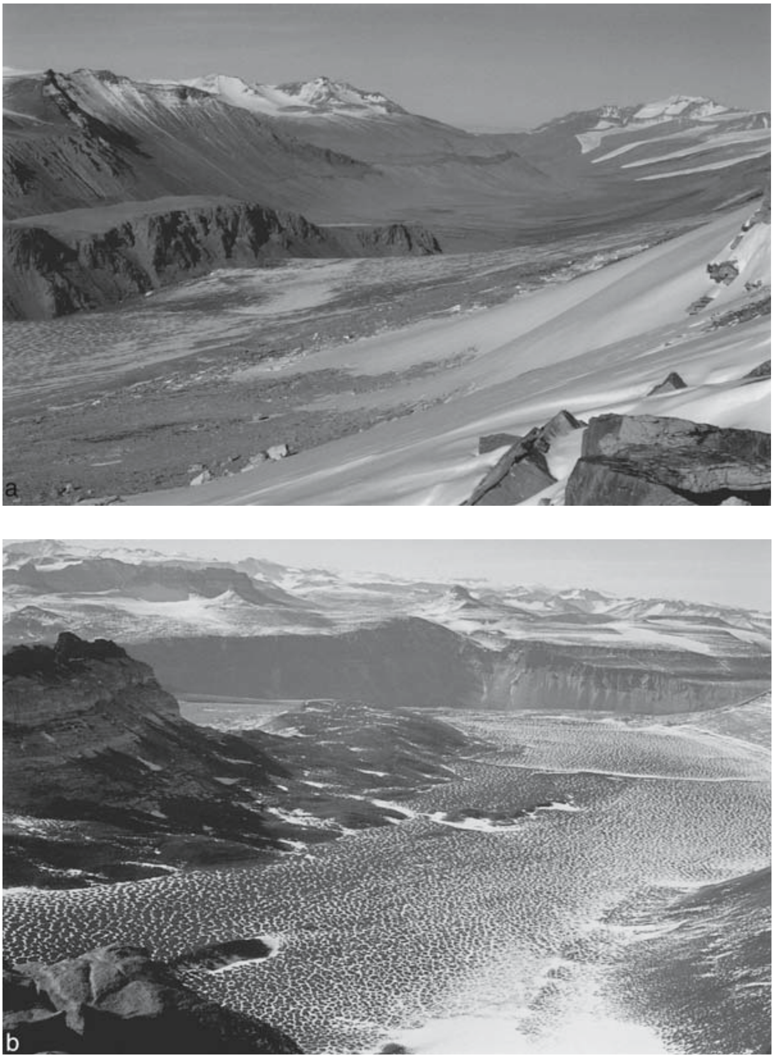

Fig. 2.3 **a** View looking east towards the coast along Wright Valley. Expansions of the Ross Ice Shelf deposited moraines in the valley mouth with earlier incursions extending far up the valley. The four alpine glaciers on the far right have moraine sequences dating to >2.1 million years. Foreground surfaces have old Miocene aged tills and soils. Wright Valley, formerly a fjord, was probably carved by a through-flowing glacier in the Oligocene. **b** View looking northwards along the Transantarctic Mountains and across Wright Valley. Tills with patterned ground are alpine moraines with weathering stage 2 soils. An older landscape and soils occur on the rounded patterned ground-free terrain in the middle

Transantarctic Mountains (Denton and Hughes 1981). The last expansion in the Late Last Glacial period (Ross Glaciation; Denton et al. 1971) resulted in widespread deposition of tills to more than 1,000 m elevation in valleys and coastal surfaces. Alpine glaciers are small and independent of the ice sheets, and comprise ice from snow accumulations in local névés, etc. These glaciers respond to changes in local conditions and, like the East Antarctic Ice Sheet, have moraine sequences which indicate that changes in their masses since the late Pliocene have been relatively small (Everett 1971).

Tills that are associated with the three ice sources, and in which the soils and permafrost occur, have broadly similar characteristics, usually diamictons which are predominantly bouldery sands or silty sands. Tills on older inland surfaces are mainly unconsolidated, often deeply weathered and sometimes include several layers separated by paleosols, which are indicative of multiple ice advances. Younger tills, especially those of the Ross Glaciation, are typically unweathered and firmly ice-cemented, while some tills are underlain by massive ice that is believed to be several million years old (Campbell and Claridge 1987; Sugden et al. 1999). Tills cover most of the exposed landscapes throughout Antarctica, but steep slopes, upland plateau and benched surfaces commonly have bedrock outcrops and felsenmeer that are estimated to make up 10–15% of all bare ground surfaces. Aeolian deposits and fluvial deposits are rarely found.

2.4 The Biological Environment

The Antarctic soil biological environment is known from many studies including those of Gressitt (1967), Cameron (1971), Holdgate (1977), Friedmann (1982), Broady (1996), Powers et al. (1995), Vishniac (1996) and Green et al. (1999); see also Chaps. 9–12 in this book. The terrestrial biota have a sporadic occurrence, being found only in very small areas where there is sufficient light, water, warmth and shelter from wind. Biodiversity is extremely low, and diminishes with increasing severity of climatic conditions. Primary producers are bryophytes, lichens, cyanobacteria and algae, and terrestrial fauna include collembola, mites and groups of microscopic organisms. In the warmer Antarctic Peninsula and other maritime areas, lichen, moss and vascular plants form communities that may give rise to peat formation, with soils that are modified by incorporation of organic matter (Blume et al. 1997). Elsewhere, and also apart from penguin nesting areas, there is no organic influence on the soils.

2.5 Physical Properties of Antarctic Soils and Permafrost

The physical properties of Antarctic soils and permafrost are known from numerous studies since the 1960s, but principally from those of Ugolini (1964), Claridge (1965), Campbell and Claridge (1975, 1987, 2006), Claridge and Campbell (1977), Bockheim (1979), Blume et al. (1997) and Campbell et al. (1998). The two main pedological processes that operate in Antarctic soils are oxidation and salinization. Coarse particle reduction takes place mainly at the soil surface, with surface clast size becoming smaller through granular disintegration and abrasion. Within the soil, coarse particles are nearly always angular and unstained, indicating low cryoturbic activity. The organic regime is everywhere insignificant, owing to the paucity of biological communities.

2.5.1 Principal Soil Weathering Processes

Oxidation, or reddening of the soil, derives from the very slow oxidation of iron-bearing minerals in rock particles, and usually results in a thin coating of iron oxides on mineral grains. The youngest soils have colours resembling those of rock, but as soil age increases the intensity of oxidation and reddening and the depth of oxidation both increase, with alteration extending to beyond 1 m in depth. Salinization, or the accumulation of salts, is widespread, and is a consequence of high evaporation rates, which typically exceed precipitation. The salts in the soils may form distinct horizons, and are predominantly derived from atmospheric transport. Clear geographic and climate-related differences in soil salt content, as well as age-related differences in salt abundance, are found (Claridge and Campbell 1977; Campbell and Claridge 1987). Salt accumulation is essentially linear with time (Bockheim 1979), and chemical weathering is insignificant by comparison.

2.5.2 Soil Morphological Properties

Antarctic soils are coarse-textured, with coarse particles >2 mm typically exceeding 50% (Table 2.1). Horizon development is weak, and mostly restricted to colour changes that diminish in intensity with increasing depth, to lithologically related textural changes, or to the presence of salt accumulations (Fig. 2.4). The soil surface is usually a stone pavement including loose material derived from fragmentation of surface clasts. On younger surfaces, clasts are mainly angular, coarse and unweathered, while on older surfaces, clast rounding, rock pitting, ventifaction, oxidation and disaggregation may be prominent. Weakly developed vesicular structure may be present in the surface horizon as a result of freezing when the soil is moist. Where there is an increased proportion of fine material, a thin surface crust may be present. Below the surface, the soil is usually structureless and pulverulent, except where salt concentrations occur, when the soil material may be firmly cohesive. In older soils, the disaggregation of coarse-grained clasts by salt weathering results in rock ghosts that indicate a highly stable soil environment.

Table 2.1 Coarse fraction (weight %) for a typical soil from Marble Point, McMurdo Dry Valleys area

	Weight (%) of coarse fractions				
Soil depth (cm)	2–5 mm	5–20 mm	20–75 mm	0.1–75 mm	>2 mm (whole soil)
0–3	7	55	4	90	66
3–15	6	11	41	88	58
15–32	7	11	39	87	57
32–45	8	12	53	92	73
45–69	11	22	41	91	74
69–100	10	20	79	79	49

Fig. 2.4 Profile of weathering stage 3 soils from dolerite and sandstone till from the Asgard Range in Wright Valley. The surface pavement is well-developed, with moderate reduction, rounding and staining of surface boulders. A weakly developed salt horizon is present, with a concentration of salts to the right of the tape beneath a boulder that was removed. Ice-cemented permafrost is at 35 cm

2.5.3 Soil Distribution Patterns

With increasing time, soil oxidation intensity and oxidation depth, as well as the soil salt content, increase. Campbell and Claridge (1975) found that soil weathering indicated by these parameters could be expressed in terms of six soil weathering stages

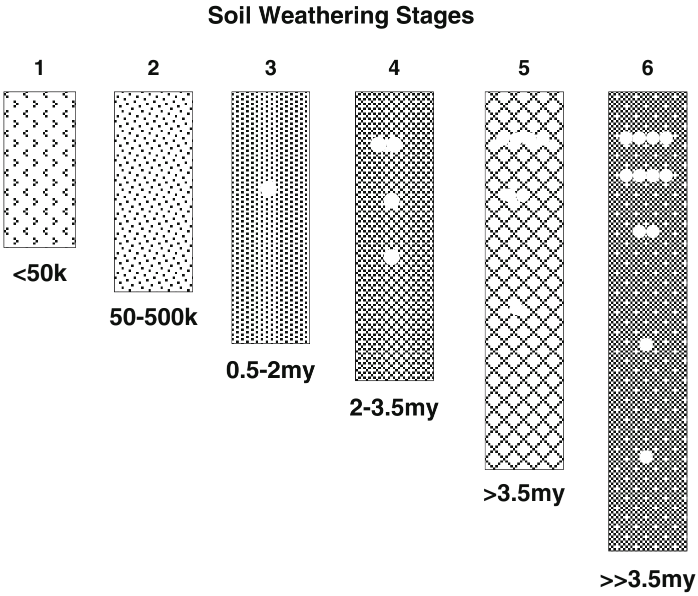

Fig. 2.5 Weathering stages identified in Antarctic soils by Campbell and Claridge (1975) are marked by increasing intensity and depth of oxidation and increasing soluble salt content. Weathering stage 5 soils may be Miocene or older judged by subsequent dating of volcanic ashes. *k* = 1,000 years; *my* = million years

covering the time between late Last Glaciation and the Miocene (Fig. 2.5). These weathering differences are intimately associated with landform differences, most commonly moraine sequences of differing ages. Coupled with the soil age differences are soil differences resulting from climate. Soils in the oceanic subxerous and moist zones, for example, have comparatively high water contents, grading from around 0.5% in surface horizons to 12% near the permafrost boundary, while soils in the arid ultraxerous zones may have a moisture content of <0.5% through the whole profile. The soil salt content likewise shows a marked geographic distribution pattern, the coastal soils having salts dominated by sodium chloride, and the arid inland soils by nitrate salts.

2.5.4 Antarctic Permafrost Properties

Antarctic soils are everywhere underlain by permafrost, which can be divided into a number of distinct types (Campbell and Claridge 2006). Ice-cemented or ice-bonded permafrost (Fig. 2.6) is easily recognized, and has an active layer that immediately overlies hard ice-bonded permafrost. The active layer depth varies according to mean annual temperature, moisture supply and the thermal radiation balance, but is usually deepest (up to 1 m) in warmer northern locations, and shallow (<2 cm) in the coldest areas. A similar form is permafrost with massive ice

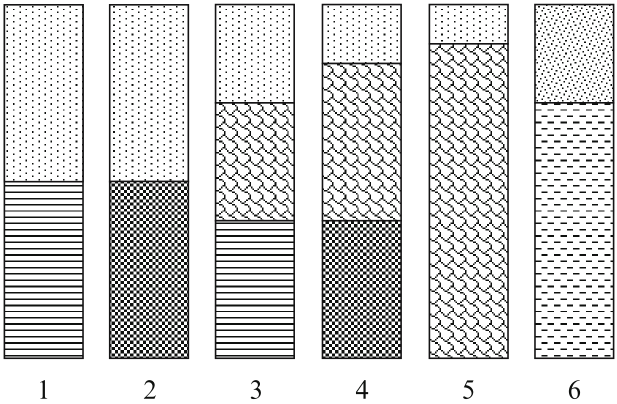

Fig. 2.6 Permafrost types in the Transantarctic Mountains region. The active layer thickness diminishes with increasing coldness and with increasing age and aridity, the permafrost changes from ice bonded to dry permafrost. *1*: active layer over ice-bonded permafrost, *2*: active layer over buried or massive ice, *3*: active layer over dry permafrost over ice-bonded permafrost, *4*: active layer over dry permafrost over buried or massive ice, *5*: active layer over dry permafrost, *6*: saline permafrost

immediately below or at some depth below the active layer. This ice is typically stagnant or old residual glacial ice (Claridge and Campbell 1968; Sugden et al. 1993), commonly associated with patterned ground surfaces (Fig. 2.4) and younger land surfaces with thermokarst terrain (Campbell and Claridge 2003).

Ice-free or dry permafrost (Bockheim 1995) is distinguished by very low water content in both the active layer and the permafrost, which is loose and non-cohesive. Ice crystals, where present, may behave like sand grains. Our measurements indicate that a gravimetric water content of around 6–7% is required for ice bonding to occur in these sandy gravel materials. In ice-bonded permafrost, weathering is restricted to the active layer but in dry permafrost, weathering occurs into the permafrost, sometimes to a depth of several meters. Intermediate forms between ice-bonded and dry permafrost are also found with dry and weathered perennially frozen permafrost overlying at variable depth ice-bonded permafrost or ancient massive ice (Claridge and Campbell 1968; Sugden et al. 1999).

Saline permafrost is found in small depressions and salty hollows, and associated soils are highly saline. In summer months, the active layer frequently contains brine, usually at a temperature several degrees below 0°C, while the soil is characterized by abundant efflorescences of soluble salts.

2.5.5 Permafrost Distribution Patterns

The distribution of the differing permafrost types, based on more than 900 observations from northern Victoria Land and through the Transantarctic Mountains, was summarized by Campbell and Claridge (2006). The permafrost table is at greatest depth in the warmer northern regions of Antarctica, and diminishes in depth with increasing latitude and altitude, with some soils possibly being perennially frozen. There is much site variation, however, due to local differences in the heating from radiation owing to topographic shading, aspect, snow cover, surface colour and surface roughness.

Ice-bonded permafrost is most commonly found in coastal regions, on younger-aged surfaces nearest to a glacier and in areas where the precipitation or drainage regime results in moist soils. At higher elevations and greater distances inland, on the older land surfaces and areas of greatest aridity such as parts of the Dry Valleys, dry permafrost, including the intermediate form, predominates. When the transition from one form to another occurs over a short distance, it is commonly related to surface age or moisture availability differences. The ice content of ice-bonded permafrost is usually greatest in coastal regions and least in colder regions. Permafrost is also present in exposed bedrock surfaces, where it may be either ice-bonded or dry.

2.6 Chemical Properties of Antarctic Permafrost Soils

Chemical weathering is of very low intensity in these soils, because of the low temperatures and extreme aridity, but soils vary in their chemistry because of environmental variations. The soils contain very small amounts of fine particle size material and even less of clay-sized material, which is the most chemically reactive fraction of the soil. Most of the fine particle size material is produced by physical disintegration of the Beacon Supergroup sandstones, so that the fine fraction of the soil is dominated by rounded quartz grains of fine sand grade, together with smaller amounts of material produced by glacial grinding.

Clay-sized material largely originates from the matrix bonding the sandstones together, and consists of micas and vermiculites of little chemical reactivity. In some instances, these have been altered by soil weathering processes to more hydrous clay minerals, illites, hydrated vermiculites and (in rare instances) smectites. In some old soils, especially those of higher weathering stages, authigenic clay minerals may be formed. The nature of these minerals is dependent on factors such as soil pH and the chemistry of the salts,

Because the climate is extremely arid, salts, mainly derived from precipitation, accumulate in the soils and strongly influence the soil chemistry. Near the coast, where winds from the sea may carry ocean-derived salts some distance inland, the soil salts are largely chlorides and sulphates of sodium, and the soils are alkaline — up to pH 9 in some cases. This may cause the transformation by hydration of some of the micas into illites and more hydrous clays, even forming some smectites (Claridge 1965). Because the buffering capacity of the soil is very low, only small amounts of salts are needed to raise the pH of the soil to high levels. Soils close to the coast are also generally very young, and contain comparatively low amounts of salts.

Further inland, soils are older, and salts have accumulated to a much greater extent than in coastal regions, often forming thick salt horizons. The salts in these soils are considered to have been derived from the oxidation of protein material caught up from the ocean surface and transported through the upper atmosphere, where they become completely oxidised to nitric acid and sulphuric acid (Claridge and Campbell 1977). Other mechanisms are also proposed, such as auroral fixation of nitrogen. Soils of inland regions therefore have low pH values; as low as 6.0 in ultraxerous soils of weathering stage 5 on the inland edge of the Transantarctic Mountains. The pH of the soil can be directly related to distance from the open sea.

Some breakdown of primary minerals takes place in the acid environment of these soils. The ferromagnesians in particular release iron, which causes the reddish staining on grain surfaces as the iron is oxidised in older soils. Cations such as calcium and magnesium are released, so that the soluble salts, which are such a dominant feature of the older soils of inland regions, are nitrates and sulphates of calcium and magnesium. Almost all crystalline phases that can be formed by combinations of calcium, magnesium, sodium, nitrate and sulphate can be identified in the soils (Claridge and Campbell 1977).

Because the salts in solution lower the freezing point, liquid water can be present at very low temperatures, generally as thin films on grain surfaces, and chemical processes can take place at temperatures as low as –50°C. In most of the old, weathering stage 5 soils of the inland edge of the Transantarctic Mountains, the clay-sized fraction of soils formed on till is dominated by clays derived from Beacon Supergroup rocks. However, in some soils formed directly on physically fragmented dolerite, these clays are absent, and it is possible to demonstrate the formation of authigenic clays, such as nontronitic montmorillonite, a consequence of clay mineral formation in an environment rich in iron and magnesium (Claridge and Campbell 1984). In these cases, though, most of the clay-sized material is physically disintegrated fragments of the glassy matrix of the parent rock of the soil, Ferrar Dolerite. In some situations, especially the very old soils of the inland regions, zeolites such as chabazite (Dickinson and Grapes 1997) may form.

Thus, the chemistry of the soil depends on geographic location, which determines the nature of the salts, the weathering processes operating and the secondary mineral that may be formed.

2.7 Sensitivity to Change

Because weathering processes in Antarctica are infinitely slow, terrestrial ecosystems in this harsh environment are extremely fragile. A wide-ranging review of the impacts of human activitie*s* and the susceptibility of the land systems to disturbance was carried out for the Ross Sea region (Campbell 2001), and showed that disturbances from human activities are long-lasting. Physical disturbances to the soils may persist for many hundreds of years, or in the most arid zones where recovery processes are negligible, be permanent. Chemical contaminations may also persist in the absence of significant leaching. Permafrost is likewise dramatically and rapidly altered when physical disturbance takes place. Less clear, however, are the future impacts of global climate change. Over recent decades, a distinct warming trend has been noted in the Antarctic Peninsula region, while recent data suggests that there may be a cooling trend in the East Antarctic region.

2.8 Conclusion

The soils of Antarctica are for the most part formed in the absence of biological processes and, as a consequence of the prevailing low temperatures, are everywhere underlain by permafrost, with the active layer varying in thickness from about one metre in northern areas to a few centimeters or less in the soils of the inland edge of the Transantarctic Mountains. The permafrost is generally ice-cemented, but in older and drier soils may be loose. Because of the extreme aridity, the soils accumulate salts derived from precipitation and weathering, the composition and amount of the salts being a function of soil age, composition of the parent material and distance from the coast. Chemical weathering processes are assisted by the salts, which allow unfrozen saline solutions to be present on grain surfaces and cracks in rock particles, even at very low temperatures. Weathering comprises the breakdown of ferreomagnesian minerals, releasing iron and cations to the soil solution. The iron oxidises and is precipitated on grain surfaces, giving rise to the red colouring of older soils. The cations, especially calcium and magnesium, combine with nitric and sulphuric acids arriving in precipitation, to make up part of the thick salt horizons which are found in older soils. The concentrated salt solutions react with silica, also released by weathering, to form secondary clay minerals and in some cases, zeolites.

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