Literature review of glaciers monitoring from remote sensing techniques

Seminar Final Report

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Glacier and ice fields\textsuperscript{1} are recognize as sensitive indicators of climate change and their importance is specially significant because they contain a considerable part of the world’s fresh water. They are also a critical component of the Earth’s system considering that the current accelerated melting and retreat of glaciers have severe impacts on the environment and human well-being, such as natural disasters, water supply, vegetation patterns, economic livelihoods, and specially they can contribute to sea level change.

As global extent of snow has a great influence on the Earth’s climate, parameters such as extent, volume, depth, etc., are crucial for many applications, like runoff assessment, water management and flood control. Hence, the study of parameters like albedo, snow grain size and surface temperature are applicable also to snow cover on glaciers and ice sheets, which are key input parameters for various models in the climate monitoring field.

A glacier is a large mass of ice having its genesis on land and represents a multi-year surplus of snowfall over snow melt. At the present time, perennial ice covers about 10\% of the land areas of the Earth.

Although glaciers are generally thought of as polar entities, they also are found in mountainous areas throughout the world, on all continents, and even near the Equator, on high mountains in Africa and South America.

In the field of glaciology, satellite remote sensing became a convenient tool for studding glaciers and ice sheets, allowing to conduct research to over large and otherwise inaccessible areas.

\textsuperscript{1}This report is mainly oriented to develop theory and applications related to the concept of glacier revealed in section 2.1, even though \textit{ice fields} and \textit{ice sheets} are also considered because they are sources of origin of many glacial tongues, called \textit{outlet glaciers}. 
Further, in high latitude regions like Greenland and Antarctic, it is only during parts of the year that effective ground-based research can be carried out due to the harsh climate environment and the lack of daylight. Satellite remote sensing often permits real-time, year-round and long-term studies.

Satellite data help in understanding process on the regional and global scale, for example, global satellite-derived maps of snow cover. Such products are particularly important because they assist interpretation and analysis concerning global change. Also, on smaller scales, satellite remote sensing can be a vital tool both for obtaining a basic understanding of processes, e.g., glacier velocities from interferometry products and glacier retreat detection for climate change monitoring programs.

Building upon different studies, this report propose a literature review about glacier monitoring from remote sensors, like optical and SAR (Synthetic Aperture Radar) satellites, including some specific applications. The review is mainly oriented to deep in the state of art of glacier monitoring. Additionally, it provides complementary information referred to actual monitoring programs and international efforts in this topic, recognized (as said before) as sensitive indicator of global climate change.

The work is organized in six chapters, comprising at first a short introduction. Chapter 2 contains a description about current review of the topic, including fundamental concepts to understand glacier dynamic and a current review about glacier situations at different scales (global, regional and local). Chapter 3 exposes a review of spectral behavior of glaciers, specifically in the optical, infrared and microwaves region of the electromagnetic spectrum. Chapter 4 contains the main applications in the remote sensing field to monitoring glaciers (although exists many others not developed here), including passive remote sensing data and active sensors, the latter specially dedicated to detect small changes in mass and structure. Subsequently, Chapter 5 gives a brief review of remote sensing techniques available for environmental hazards related to glaciers. Finally, the conclusions is presented in Chapter 6.
CHAPTER 2

General concepts and current overview

2.1 What is a glacier?

A glacier is a large mass of ice having its origin on land and normally displaying some movement in response to gravity or its own weight [1] [2]. The input to a glacier is in the form of snowfall, and the output is principally in the form of melt water and, in the case of a glacier whose lower terminus is in water, icebergs can calve off and float away. This determine an input area of accumulation and a output area named of ablation, both separated by the equilibrium line.

The surface of a glacier can be divided into a number of zones or facies (illustrated in Figure 2.1).

1. Uppermost is the dry-snow zone in which no melting takes place. This zone is only found inland in Greenland and over most of the Antarctic ice sheet, and on the highest mountain glaciers, where the annual average temperature is lower than around 11°C.

2. Next is the percolation zone, in which some surface melting occurs during the summer. The meltwater percolates downward and refreezes to form inclusions of ice in the form of layers, lenses, and pipes. The dry-snow and percolation zones are separated by the dry-snow line.

3. Below the percolation zone, and separated from it by the wet-snow line, is the wet-snow zone in which all the current year’s snowfall melts.

4. Below this, and separated from it by the snow line (sometimes called the firn line) is the superimposed ice zone. In this zone, surface melting is so extensive that the
meltwater refreezes into a continuous mass of ice. The lower boundary of this zone is the equilibrium line.

Not all of these zones are present in all glaciers because only the coldest glaciers possess dry-snow zones. Temperate glaciers, in which the temperature of all but the upper few meters is at the freezing point, exhibit only wet-snow and ablation zones (the superimposed ice zone is generally negligible so the snow line and equilibrium line coincide). Ice shelves, which are parts of glaciers or ice sheets that extend over water, do not have ablation zones but instead lose material mostly by calving of glaciers [1].

Figure 2.1: Schematic diagram of an idealized glacier [3].

Snow is transformed to ice in a glacier by a variety of mechanisms, including mechanical settling, sintering (in the dry-snow zone), refreezing of meltwater, and refreezing of sublimated ice to form depth hoar. The density of the material in a glacier increases with depth and once the transformation process has begun, the material is referred to as firm (see Appendix A) rather than as snow, and firm (also called nevé) generally has a density greater than 0.55Mgm$^{-3}$. Firm is porous, since it contains interconnected air channels. However, once the density increases above about 0.83Mgm$^{-3}$ these channels are closed off, resulting in ice in which closed air bubbles are trapped [1].

2.2 Glaciers and climate

Glacier is one of the elements of the system called Cryosphere. Other constituents are the snow cover, sea ice, freshwater ice (frozen lakes and rivers), the large ice masses on land (glaciers and ice sheets, and related phenomena such as ice shelves and icebergs), and permafrost.
Temperatures on the Earth’s surface generally decrease with increasing distance from the equator, so the cryosphere is largely a high-latitude phenomenon [1]. When referring to perennial surface ice on land, one usually differentiates between ice sheets, ice shelves, glaciers and ice caps (Figure 2.2).

There are fundamental differences in time-scales and processes involved between the different components of the perennial surface-ice on land. Due to the large volumes and areas, the two continental ice sheets actively influence the global climate over time scales of months to millennia. Glaciers and ice caps, with their smaller volumes and areas, react to climatic forcing at typical time scales from years to centuries [4].

Climate change is now clearly at the top of the world’s agenda. Glaciers and ice caps are now also one of the Essential Climate Variables, a set of core variables in support of the work of organizations such as the United Nations Framework Convention on Climate Change (UNFCCC) and the Intergovernmental Panel on Climate Change (IPCC) [4].

Changes in glaciers and ice caps provide some of the clearest evidence of climate change, and as such they constitute key variables for early detection strategies in global climate-related observations. These changes have impacts on global sea level fluctuations, the regional to local natural hazard situation, as well as on societies dependent on glacier meltwater.

Glaciers, ice caps and continental ice sheets cover some 10% of the earth’s land surface at the present time, whereas during the ice ages, they covered about three times this amount. The present ice cover corresponds to about three-quarter of the world’s total freshwater resources. If all land ice melted away, the sea level would rise by almost 65 m, with the ice sheets of Antarctica and Greenland contributing about 57 and 7 metres, respectively, and all other glaciers and ice caps roughly half a metre to this rise [4].

In the current scenarios of climate change, the ongoing trend of worldwide and rapid, if not accelerating, glacier shrinkage on the century time scale is most likely of a non-
periodic nature, and may lead to the deglaciation of large parts of many mountain ranges in the coming decades. Such rapid environmental changes require that the international glacier monitoring efforts make use of the swiftly developing new technologies, such as remote sensing, and relate them to the more traditional field observations [4].

2.2.1 Observations of climate change

"Warming of the climate system is unequivocal, as is now evident from observations of increases in global average air and ocean temperatures, widespread melting of snow and ice and rising global average sea level" [5].

Eleven of the last twelve years (1995-2006) rank among the twelve warmest years in the instrumental record of global surface temperature (since 1850). The temperature increase is widespread over the globe and is greater at higher northern latitudes. Average Arctic temperatures have increased at almost twice the global average rate in the past 100 years. Land regions have warmed faster than the oceans. Observations since 1961 show that the average temperature of the global ocean has increased to depths of at least 3000 m and that the ocean has been taking up over 80% of the heat being added to the climate system. The sea level rise agreed with this warming. Average, the world level ocean has increased 1.8 mm/yer from 1961, and 3.1 mm/yer from 1993, cause explained by the melting of glaciers, ice sheets and polar caps [5].

Observed decreases in snow and ice extent are also consistent with warming. Satellite data since 1978 show that annual average Arctic sea ice extent has shrunk by 2.7 [2.1 to 3.3]% per decade, with larger decreases in summer of 7.4 [5.0 to 9.8]% per decade. Mountain glaciers and snow cover on average have declined in both hemispheres. The maximum areal extent of seasonally frozen ground has decreased by about 7% in the Northern Hemisphere since 1900, with decreases in spring of up to 15%. Temperatures at the top of the permafrost layer have generally increased since the 1980’s in the Arctic by up to 3°C (Figure 2.3).
At continental, regional and ocean basin scales, numerous long-term changes in other aspects of climate have also been observed. Trends from 1900 to 2005 have been observed in precipitation amount in many large regions. Over this period, precipitation increased significantly in eastern parts of North and South America, northern Europe and northern and central Asia whereas precipitation declined in the Sahel, the Mediterranean, southern Africa and parts of southern Asia. Globally, the area affected by drought has likely increased since the 1970’s. Some extreme weather events have changed in frequency or intensity over the last 50 years:

- It is very likely that cold days, cold nights and frosts have become less frequent over most land areas, while hot days and hot nights have become more frequent.
- It is likely that heat waves have become more frequent over most land areas.
- It is likely that the frequency of heavy precipitation events (or proportion of total rainfall from heavy falls) has increased over most areas.
- It is likely that the incidence of extreme high sea level has increased at a broad range of sites worldwide since 1975 [5].
There is observational evidence of an increase in intense tropical cyclone activity in the North Atlantic since about 1970, and suggestions of increased intense tropical cyclone activity in some other regions where concerns over data quality are greater. Multi-decadal variability and the quality of the tropical cyclone records prior to routine satellite observations in about 1970 complicate the detection of long-term trends in tropical cyclone activity.

Average Northern Hemisphere temperatures during the second half of the 20th century were very likely higher than during any other 50-year period in the last 500 years and likely the highest in at least the past 1300 years.

The statements presented here are based largely on data sets that cover the period since 1970. The number of studies of observed trends in the physical and biological environment and their relationship to regional climate changes has increased greatly since 2001. The quality of the data sets has also improved. There is a notable lack of geographic balance in data and literature on observed changes, with marked scarcity in developing countries. These studies have allowed a broader and more confident assessment of the relationship between observed warming and impacts than was made in 2001. That assessment concluded that "there is high confidence that recent regional changes in temperature have had discernible impacts on physical and biological systems" [5].

Observational evidence from all continents and most oceans shows that many natural systems are being affected by regional climate changes, particularly temperature increases. There is high confidence that natural systems related to snow, ice and frozen ground (including permafrost) are affected. Examples are:

- enlargement and increased numbers of glacial lakes.
- increasing ground instability in permafrost regions and rock avalanches in mountain regions.
- changes in some Arctic and Antarctic ecosystems, including those in sea-ice biomes, and predators at high levels of the food web

Based on growing evidence, there is high confidence that the following effects on hydrological systems are occurring: increased runoff and earlier spring peak discharge in many glacier- and snowfed rivers, and warming of lakes and rivers in many regions, with effects on thermal structure and water quality [5].

### 2.2.2 International efforts for glaciers monitoring

Worldwide information about ongoing glacier changes was initiated in 1894 with the foundation of the Commission Internationale des Glaciers at the 6th International Geological Congress in Zurich, Switzerland. A first attempt to compile a world glacier inventory started in the 1970’s based mainly on aerial photographs and maps. Up to now, it resulted in a detailed inventory of more than 100,000 glaciers covering an area of about 240.000 km$^2$, and in preliminary estimates for the remaining ice cover of some 445.000 km$^2$. Today
the task of inventorying glaciers worldwide is continued for the most part based on satellite images [4].

Nowadays, the World Glacier Monitoring Service (WGMS) continues the collection and publication of standardized information on distribution and ongoing changes in glaciers and ice caps. The WGMS is a service of the International Association of the Cryospheric Sciences of the International Union of Geodesy and Geophysics (IACS, IUGG) and the Federation of Astronomical and Geophysical Data Analysis Services of the International Council for Science (FAGS, ICSU) and maintains a network of local investigators and national correspondents in all the countries involved in glacier monitoring. In cooperation with the US National Snow and Ice Data Center (NSIDC) in Boulder and the Global Land Ice Measurements from Space (GLIMS) initiative, the WGMS is in charge of the Global Terrestrial Network for Glaciers (GTN-G) within the Global Climate/Terrestrial Observing System (GCOS/GTOS). GTN-G aims to combine (a) field observations with remotely sensed data, (b) process understanding with global coverage, and (c) traditional measurements with new technologies by using an integrated and multi-level monitoring strategy [4].

The need for a worldwide inventory of existing perennial ice and snow masses was first considered during the International Hydrological Decade declared by UNESCO for the period of 1965 to 1974. These tasks were continued by its successor organization, the WGMS, after 1986. In 1989, a status report on the WGI was published including detailed information on about 67,000 glaciers covering some 180,000 km$^2$ and preliminary estimates for the other glacierised regions, both based on aerial photographs, maps, and satellite images (WGMS 1989). The detailed inventory includes tabular information about geographic location, area, length, orientation, elevation and classification of morphological type and moraines, which are related to the geographical coordinates of glacier label points [4].

2.3 Global distribution of Glaciers and Ice caps

Distribution of glaciers is described in Figure 2.4, which provide an overview on the glacier changes after the Little Ice Age (LIA) in eleven glacierised macroregions. The eleventh macroregions are classified according to the extent of the glacier cover.

The LIA was a cold period that occurred from the early 16th to the mid 19th centuries, after the so-called Medieval Warm Period. Possible causes of the Little Ice Age are orbital cycles, decreased solar activity, increased volcanic activity, altered ocean current flows and the inherent variability of global climate. It had three particularly cold intervals: one beginning about 1650, another about 1770, and the last in 1850, each separated by intervals of slight warming. It could be said that it ended around 1850 [6].

In the following lines a review of global distribution of glaciers and ice sheets (following to [4]) will be done, according to the eleven microregions defined.
2.3 Global distribution of glaciers and ice caps

1. **New Guinea.** The only tropical glaciers of Asia are located on the mountains of New Guinea. The few glaciers of Papua (formerly Irian Jaya, Indonesia) and Papua New Guinea are located on the peaks of the great Cordillera of the island of New Guinea. Most observations focused on the glaciers have undergone extensive retreat since the LIA maximum extent reducing the entire Puncak Jaya ice cover from almost 20 km\(^2\) around 1850 to less than 3 km\(^2\) in 2002, with highest retreat rates around 1940 and in the early 1970s. All ice masses except some on Puncak Jaya have now disappeared.

2. **Africa.** The few tropical ice bodies in East Africa are located on Ruwenzori, Mount Kenya and Kilimanjaro. African glaciers are found near the equator in East Africa, situated on three mountains: Ruwenzori (5,109 m asl), Mount Kenya (5,199 m asl) and Kilimanjaro (5,895 m asl), of which the latter are volcanoes. The ice cover on Ruwenzori has retreated continuously since the late 19th century, became strongly fragmented and on some peaks has completely vanished. The ice bodies on Kilimanjaro have shrunk continuously from about 20 km\(^2\) just before 1880 to about 2.5 km\(^2\) in 2003. The ice volume of Lewis Glacier (Mount Kenya) decreased from about 7.7 km\(^3\) in 1978 to about 0.3 km\(^3\) in 2004 with an average thickness loss of almost one metre ice per year.

3. **New Zealand.** Most glaciers are situated along the Southern Alps, with a few more on Mount Ruapehu Volcano on the Northern Island. New Zealand’s glaciers lost between one quarter and almost half of their area between the timing of their LIA maximum extents and the 1970’s. After the mid 1980’s many glaciers on the west coast have gained mass and advanced noticeably. Since the beginning of the 21st century, the number of retreating glaciers has increased again.

4. **Scandinavian.** The majority of the ice on the Scandinavian Peninsula is located in southern Norway. Some glaciers and ice caps are also found in northern Norway and the Swedish Kebnekaise mountains. Due to the combination of high latitude and the moisture from the North Atlantic, many glaciers and ice caps developed, mainly in Norway, all within 180 km of the west coast. Glaciers in Scandinavia have an important role because nowadays, 15% of the used runoff comes from glacierised basins and 98 per cent of the
2.3 Global Distribution of Glaciers and Ice Caps

electricity is generated by hydropower production. In general, Scandinavian glaciers experienced a recession during the 20th century with intermittent periods of re-advances around 1910 and 1930, in the second half of the 1970s, and around 1990; the last advance stopped at the beginning of the 21st century.

5. Central Europe. Glaciers are found in the European Alps, the Pyrenees, and the Caucasus Mountains. The Alps represent the water tower of Europe and form the watershed of the Mediterranean Sea, the North Sea/North Atlantic Ocean, and the Black Sea. The front variations show a general trend of glacier retreat over the past 150 years with intermittent Alpine glacier re-advances in the 1890’s, 1920’s, and 1970/1980’s. The Alpine glacier cover is estimated to have diminished by about 35% from 1850 to the 1970’s and another 22% by 2000. Mass balance measurements show an accelerated ice loss after 1980 culminating in an annual loss of 5 to 10% of the remaining ice volume in the extraordinarily warm year of 2003. In the Caucasus, glacier retreat since the end of the LIA is also widespread, with a certain amount of mass gain in the late 1980s and the early years of the 21st century. The recent retreat was associated with an increase in debris cover and glacier lake development. Since the first half of the 19th century, about two thirds of the ice cover was lost in the Pyrenees with a marked glacier shrinking after 1980.

6. South America. Glaciers are widespread along the Andes from the tropical ice bodies in the north to the Patagonian Icefields and the Tierra del Fuego in the south. Approximate glacier areas for tropical South America are: 1.8 km$^2$ for Venezuela, 87 km$^2$ for Colombia, 90 km$^2$ for Ecuador, 1780 km$^2$ for Peru and 534 km$^2$ for Bolivia. By far the largest ice cover at about 23,000 km$^2$ is found in Chile and Argentina, with more than 85% located in the Northern and Southern Patagonian Icefields and in the Cordillera Darwin Icefield in Tierra del Fuego. Situation of glaciers in South America will be developed in section 2.4.

7. Northern Asia. The majority of land surface ice in Northern Asia is located on the East Arctic Islands such as Novaya Zemlya, Severnaya Zemlya and Franz Josef Land, as well as distributed in the mountain ranges from the Ural to the Altay, in the east Siberian mountains and Kamchatka. On average, the scale of glacier shrinkage was much smaller in continental Siberia than in central Asia and along the Pacific margins. On Kamchatka both retreats and advances have occurred on glaciers influenced by volcanoes, whereas a general retreat was found on glaciers located in the coastal area.

8. Antarctica. The vast majority of glaciers and ice caps in the Antarctica are located on the Antarctic Peninsula and around the Antarctic Ice Sheet, with an overall estimated area ranging from 70,000 km$^2$ to 169,000 km$^2$. This large uncertainty results from the difficulty to differentiate clearly between the various glaciers and ice caps, and the ice bodies closely linked to the continental ice sheet. 87% of the glaciers have retreated in the Antarctic Peninsula over the last six decades. Since 1980, most glaciers have receded; some of these retreats have been dramatic and a number of small mountain glaciers are about to disappear.

9. Central Asia. The main mountain range of Central Asia is the Himalaya and its adjacent mountain ranges such as Karakoram, Tien Shan, Kunlun Shan and Pamir. The sum of its glacierised area corresponds to about one sixth of the global ice cover of glaciers and ice caps. In general, glacier retreat was dominant in the 20th century, except for a decade
2.4 Impacts of global climate changes: The case of Argentina

or two around 1970, when some glaciers gained mass and even reacted with re-advances of a few hundred metres. After 1980 ice loss and glacier retreat was dominant again. The area loss since the 1960s is estimated to about 6 per cent, and is more pronounced in the Chinese Himalaya, Qilian Mountains and Tien Shan, but with rather small recessions in the hinterland of the Tibetan plateau.

10. North America. North American glaciers are located on mountains in the west of the continent from Alaska down to the Canadian and US Rockies, and on volcanoes in Mexico. The glacier observations show a general retreat after the LIA maximum, particularly at lower elevations and southern latitudes, which slowed down somewhat between the 1950’s and 1970’s and accelerated again after the 1970’s. Mass balance measurements show strong accelerating ice losses since the mid 1970’s which was confirmed by remote sensing studies in Alaska and Canada. In the Western Cordillera of the Rocky Mountains the glacier area loss since the LIA is estimated at about 25%.

11. Arctic islands. In the Arctic Islands, including Greenland, glaciers and ice caps are found on the Canadian Arctic Archipelago and around the Greenland Ice Sheet, as well as on the West Arctic Islands, Iceland, and Svalbard. Exists a general retreating trend of the Arctic glaciers and ice caps since the time when of their LIA extent which slowed down somewhat during the middle of the 20th century. Glaciers on Cumberland Peninsula, Baffin Island, yield an area loss of 10 to 20% between the LIA maximum extent and 2000. However, there are several regional or glacier specific variations found in this overall trend such as the mass gain of Kongsvegen (Svalbard) in the early 1990s and periods of glacier retreat (1930 to 1960, after 1990) and advance (1970 to 1985) in Iceland.

Figure 2.5 illustrate the current extension of glaciers and ice caps for each microregion.

Figure 2.5: Extension of glaciers and ice caps for macroregion. [4]

2.4 Impacts of global climate changes: The case of Argentina

Except for a few cases in Patagonia and Tierra del Fuego, glaciers in South America have shown a general retreat and wasting since the LIA maximum extent with an enhanced retreat trend in recent decades [4]. The Northern Patagonian Icefield lost about 3.4% (140
km$^2$) of its area between 1942 and 2001, whereby the frontal tongues of calving glaciers were observed to be an important source of recession and area change. Thinning rates of up to 30 m/y have been observed recently in the Southern Patagonian Icefield, with a relevant contribution to sea level rise [7].

The Northern Patagonia Icefield (NPI), located completely in Chile, and the Southern Patagonia Icefield (SPI), shared between Chile and Argentina, are the largest temperate ice masses in the Southern Hemisphere. They cover an area of 4.200 and 13.000 km$^2$ respectively; receive abundant precipitation (2 to 11 m of water equivalent per year), with a large east-west gradient; and discharge ice and meltwater to the ocean on the west side and to lakes on the east side via rapidly flowing glaciers. Few reliable mass balance data on the region exist, leaving considerable uncertainty in the estimation of its contribution to sea level rise (SLR). The fronts of most of these glaciers have been retreating over the past half century or more, and discrete measurements of thickness change in the ablation area of a few glaciers indicate rapid thinning [7].

According to [7], the Patagonia glaciers cover an area five times smaller than their Alaskan counterparts (90.000 km$^2$), yet they account for 9% of the SLR contribution from mountain glaciers versus 30% in Alaska. The contribution of Patagonia to SLR is therefore disproportionately larger (by a factor of 1.5) than is indicated by its area. They attribute this enhanced vulnerability of Patagonia glaciers to climate change to their higher turnover rates and low ELA’s (Equilibrium Line Altitude), combined with the dominance of calving glaciers.

In Argentina, glaciers and crioforms have a significant role in regional development and are emblematic components of the environmental heritage of the Andes. Among other attributes, these icy bodies are recognized as crucial components of the hydrological system of mountain and as "strategic reserves" of water to the adjacent lowlands.

Since the runoff in a basin without glaciers or crioforms depend almost exclusively on the precipitation (with extremely low or no flow during prolonged drought), to know the number, area and spatial distribution of ice bodies not only provides estimation of the water in the solid state existing in the different basins of the Andes, but also information necessary to assess the regulatory capacity of these bodies on the flows in different climatic conditions. This knowledge results essential to establish management measures and adaptation to water shortage events in different regions and different socio-economic activities that depend on water resources in arid regions of western Argentina.

The loss or drastic reduction in glacier volume observed in the Andes also carries additional risks poorly understood. The terminal moraines or deposits (formed in the past during progress of the ice front) usually enclose water when the glaciers recede, forming proglacial lakes that can increase rapidly in size. For various reasons (earthquakes, extreme rainfall, instability of deposits) these natural dams may collapse unexpectedly causing floods with potentially catastrophic consequences for human populations and infrastructure located downstream.

The case of Rio Manso Glacier flood occurred on May 21, 2009 in the area of Mount Tronador, Rio Negro Province is a typical flood caused by the collapse of a proglacial
2.4 IMPACTS OF GLOBAL CLIMATE CHANGES: THE CASE OF ARGENTINA

lake, but fortunately only affected at the bridges and infrastructure sector of Nahuel Huapi National Park.

Given the seriousness of the matter, this issue is being actively studied and monitored, especially in regions where large numbers of people living near or downstream from mountainous areas with receding glaciers and proglacial lakes formed (Peru, Himalayas). Unfortunately Argentina until this issue has not received much attention [8].

Nowadays, an inventory of glaciers and preriglacial zones is being carried out according to the sanction of the law 26.639 from October, 2010, for preservation of this environments. Its main objectives are:

- Implement appropriate methodologies for mapping and monitoring ice bodies in different regions of the country.

- Develop human resources to address the implementation and execution of the National Inventory of Glaciers and ensure its continuity in time.

- Define the type and level of detail necessary to ensure that information obtained will allow proper management of strategic reserves of water resources in Argentina.

- Organize inventory database Glacier National orderly and efficient manner using an online computer system for storage, exchange and publication of partial or final results.

- Establish an integrated system of observations "bodies of ice / climate" that would, through periodic monitoring and carefully chosen sites, identify major climatic factors affecting the evolution of strategic reserves of water resources in the short and long term.

- Laying the groundwork for continued monitoring, analysis and integration of information on glaciers and mountain range in the provinces crioformas so that national and provincial institutions to define appropriate strategies and policies for the protection, control and monitoring of its reserves water in solid state and that universities can use this information as tools for scientific research.

- Identify potential impacts of the mass loss of ice could have on water resource management and other associated human activities.

- Establish a program to disseminate information resulting from the national inventory of glaciers through an open data policy and free access of information to promote knowledge and encourage its use by public and private, policy makers, educators, scientists and the general public.
Figure 2.6: Viedma glacier in 1930 and 2008 [8].
In general terms, remote sensing can be interpreted as the gathering of information about an object without physical contact. Strictly, it refers to airborne or spaceborne observations using electromagnetic radiation. This radiation is either naturally occurring, in which case the system is said to employ passive remote sensing, or is generated by the remote sensing instrument itself (active remote sensing).

The most important types of imagery for application in glacier monitoring are optical images, but also active microwave imagery has gained an important role for this topic. The increasing in last decades of radar sensors over satellite platforms has led to scientific community to explore the use of this kind of sensors in the detection and monitoring of snow and ice parameters. High temporal frequency and high spatial resolution which is provided by active sensors, helps to increase the knowledge of its dynamics and spatial patterns.

Naturally occurring radiation includes reflected solar radiation, which is largely confined to the visible and near-infrared parts of the electromagnetic spectrum (wavelengths between roughly 0.35 and 2.5 mm) and thermally emitted radiation. The spectral region between 0.4 and 3µm is called the reflective part of the spectrum, because reflection is the predominant mechanism, whereas emission predominates in the thermal infrared region (3–100 µm).

Passive remote sensing systems that detect reflected solar radiation are designed to measure the radiance, i.e., the amount of radiation reaching the sensor in a particular waveband. If the amount of radiation that is incident on the Earth’s surface is known, the reflectance of the surface can be calculated (this requires that the effects of the atmosphere should be corrected).
The range of wavelengths generated by a thermally emitting body depends on the temperature. In the case of thermal band, the radiance reaching the instrument is measured. This radiance is normally expressed as a brightness temperature, which is the temperature of a perfect emitter (a so-called black body) that would produce the same amount of radiation. The brightness temperature is related to the physical temperature of the surface and its emissivity, a unit quantity that defines the ratio of the actual radiance to the radiance that would be emitted by a black body at the same temperature. Clearly, if the emissivity of a material is known and its brightness temperature can be measured, its actual physical temperature can be calculated [1].

Systems that operate in the microwave part of the electromagnetic spectrum are generally classed as imaging radars. The primary variable measured in this case is the backscattering coefficient, a unitless quantity related to the concept of reflectance and usually specified in decibels (dB) rather than as a simple ratio or percentage. Its dependence on the imaging geometry is often important, which can be configurated [1].

In the next sections, the behavior of glaciers in the useful regions of the electromagnetic spectrum for remote sensing will be developed.

### 3.1 Glaciers properties in the Optical and Infrared regions

The study of albedo or reflectance, snow grain size and surface temperature is applicable not only to snow cover on land, but also to snow cover on glaciers and ice sheets and can therefore also be viewed as remote sensing of glaciers.

The glacier mass can be characterized by several components as grain shape, grain size, density, hardness, temperature, impurities, layer thickness and liquid water content. Also the surface roughness plays an important role in remote sensors of microwaves, as will be explained in section 3.2.

Applications of glacier monitoring from optical, infrared and thermal remote sensing use information from the visible (VIS) part of the electromagnetic spectrum, located between 0.4 and 0.7 µm, from the near (NIR), the short wave near (SWIR) and the thermal infrared (TIR), located between 0.7 to 1.1, 1.1 to 3 and 3 to 100 µm, respectively.

In the winter, a glacier surface is usually covered by snow. The reflectance of fresh snow covering glaciers during the winter can be simply detected from optical sensors, such as Landsat TM [9], MODIS [10], MERIS, SPOT and AVIRIS [3]. However, the sensors may thus be unable to differentiate surfaces having similar reflectance in these wavelength bands. An example is the difficult differentiation between snow and clouds, which needs to be supported with specific digital processing (e.g. snow index) and field measurements. In the summer, however, other surfaces can be exposed, such as impurities in the snow cover like carbon soot, volcanic ash, and continental dust have a strong effect on reflectance, but mostly in the visible region (0.4 – 0.7 µm).

Reflectance varies strongly with wavelength, and the resulting curve is unique for each
surface type, as Figure 3.1 depicts. This analysis allows to identify various surface types and differentiate them from each other.

![Spectral reflectance curves for snow and ice in different formation stages](image1)

**Figure 3.1:** Spectral reflectance curves for snow and ice in different formation stages [1].

Snow over glaciers is not highly reflective in the thermal infrared region, but the emissivity of dry snow in the thermal infrared region ranges from typically 0.965 to 0.995. In this region of the electromagnetic spectrum the absorption of ice is high, with a maximum near 10 $\mu$m, and the finely divided structure of snow also increases its tendency to act like a black body (i.e., for the emissivity to tend toward 1). Figure 3.2 shows two recent high resolution measurements of the emissivity spectrum of snow between 3 and 15 $\mu$m. Since the emissivity of water in this range of wavelengths is not substantially different from that of snow, the effect of the presence of liquid water is negligible [1].

![Thermal infrared emissivity of two snow samples](image2)

**Figure 3.2:** Thermal infrared emissivity of two snow samples [1].
3.2 Glaciers properties in the Microwave region

As radar systems work in the microwave region of the electromagnetic spectrum, many advantages over optical imaging can be found. First, it has independent possibility of measurements of weather conditions with very low sensibility to clouds and rain. Radar is also independent of solar conditions and can measure also during night-time. The radar capabilities to penetrate surfaces and also its sensitivity to characterize physical properties of the surface, such as roughness and water content (according to the dielectric properties) are also well known. A radar system has also the capability to take very accurate distance measurements through interferometry analysis as well as the structure of the object can be detected by polarimetry analysis. At last, being an active sensor, the resolution is about 3 orders of magnitude higher than with passive microwave sensing (tens of meters versus tens of kilometers).

Radar data is considerably more difficult to interpret and understand than optical data, because many factor can affect the radar signal, such as surface roughness, the moisture state of the medium and sensor acquisition geometry.

The length of the electromagnetic wave with respect to the size of the feature determines if the surface appear rough or smooth at particular wavelength. A surface that appears rough at shorter wavelength may appear quite smooth at longer microwave wavelengths. Furthermore, incident energy direction to the antenna have to be considered. If the smooth surface is perpendicular to the antenna radar beam, then the energy returned to the radar is intense (bright areas). However, if the smooth surface is at any other direction, non or little energy is received by the antenna [2]. Most glacier surfaces appear rough at X and C frequencies, and diffuse reflection dominates over specular reflection [11].

With respect to dielectric properties, liquid water will absorb microwave radiation while snow or soil in a frozen state will allow the radar signal to penetrate.

Sensor acquisition parameters like frequency, polarization ante antenna look angle will affect also the backscattering. In general, the lower the frequency (longer wavelength) the greater will be the penetration into a medium by the radar signal. Higher frequencies will be scattered by small inclusions representing dielectric discontinuities. Additionally, if the wavelength is comparable to the size of an inclusion, the potential of volume scattering is increased. Volume scattering is the process by which the particles suspended in a medium diffuse a portion of the incident radiation in all directions, such as the case when microwave energy is scattered by crystals and grains within the glacier pack [2].

Low antenna look angles result in considerable geometric distortion and layover when radar is used in regions with high relief, like mountain areas, where most of glaciers can be found. In order to avoid "shadowing" effects, optimal antenna look angles for analysis of radar data in mountainous terrain are between 40° and 60°[2].

In general, SAR data can be difficult to interpret over glacierized terrain, especially in areas of high relief, where most of the smaller glaciers of the world are found. Nevertheless, glacier boundaries, medial and terminal moraines, undulations, snow and ice areas,
relief and roughness of glacier surfaces can be discerned easily. Medial moraines, linear feature comprised of debris that collects when two glacier flow together, can be seen clearly on SAR imagery, according to its roughness. Crevassed areas can give high values of backscatter, especially if the crevasses are oriented perpendicular to the look-direction of the radar [1]. The ablation area is also clearly distinguishable because of the abrupt decrease of the backscattering, and on the contrary, the notable increasing in the accumulation area, where layers of snow and firn contribute to volume scattering. While dry snow over ground is generally transparent at X, C and L band [11], dry snow in the accumulation areas of glaciers may provide a strong return. In the case of seasonal snow cover, the dry snow is generally not deep enough to give high returns; however, in a thick layer of snow and firn, absorption of the signal is low and the likelihood of scattering by snow grains is greater when the snow is deep, and snow grains are larger in firn versus newer snow, this results in higher returns [2, 11].
Main glacier applications using Remote Sensing data

Remote sensing of terrestrial ice masses (glaciers, ice caps and ice sheets) is generally a well-developed field. There are many approaches in glacier applications which have been developed according to the sensitivity of glaciers to climate, as was discussed in section 2.2. They include direct and indirect measurement, and approaches based in energy balance calculations for modeling, been the assessment of mass balance the major focus of research [1].

The characteristics of a terrestrial glacier or ice mass that can be measured using remote techniques include its spatial extent, surface and bottom topography, mass volume, surface flow field, accumulation and ablation rates (and hence mass balance), surface zonation, albedo, and changes in these quantities over time. As well as its importance in indicating the total amount of ice, surface topography provides important clues about the internal structure and can reveal flow features and grounding lines.

Visible and near infrared (VIR) and Synthetic Aperture Radar (SAR) imagery both play major roles in the remote sensing of terrestrial ice masses. VIR imagery, from the high albedo of snow, make it particularly easy to recognize, but as the imagery is confined to daylight and cloud free conditions, which can prove a major limitation. On the other hand, radar imagery does not suffer from these constraints, but is subject to a number of complications, such as geometric distortion and speckle [1]. Considering this, a review of the current remote sensing methods for the main glacier monitoring applications is given in the next lines.
4.1 Spatial extent and surface features

The study of glaciers extension by remote sensing methods dates back at least to the 1930’s when aerial photography was applied to the problem, and aerial photography still has considerable value. The natural extension of aerial photography to the spaceborne domain is the use of VIR imagery, and this has proved exceptionally valuable for the study of terrestrial ice masses. Many studies have been conducted by pure visual interpretation of satellite images, with little or no image processing [3].

The choice of swath width and spatial resolution in satellite imagery is governed by the spatial scale of the phenomena to be investigated, with wide swath instruments, such as AVHRR, providing a good match to the requirements of studying the large ice sheets of Antarctica and Greenland (Fig. 4.1).

Figure 4.1: AVHRR image mosaic covering the whole of Antarctica [12].

This mosaic revealed the complexity of the Antarctic ice sheet surface and provided some qualitative topographic information through shape-form-shading. Higher-resolution satellite data, such as Landsat or ASTER imagery, provide data at higher spatial resolution, showing very much more detail, including crevasses and grounding lines, supraglacial lakes, snow dunes, pitted patterns, glazed surfaces and "snow megadunes", occupying more than 500,000 km$^2$ and oriented perpendicular to the regional direction of the katabatic winds [1].

The visible bands are more suitable than infrared bands for studying topographic features. This observation is supported doing a principal components analysis with the image bands as input variables. The first principal component enhances topographic effects and is
therefore often superior to the visible bands themselves [3]. A second principal component band is mostly related to the Landsat shortwave infrared bands (TM bands 5 and 7) and enhances surfaces characteristics not seen in the visible, like patterns due to surface grain size variations [13].

Although the advantages in satellite coverage, there is a compromise between spatial resolution and spatial coverage in VIR imagery. [14] highlighted the importance of aerophotografic data to obtain high resolution as a particular interest instrument of glaciologist. In this case they compare the KFA-1000 russian camera (with a nominal resolution of 5 m.) with Landsat TM and MS and SPOT imagery (4.2), highlighting the advantage of the aerial photograph to recognize internal structure of the glacier surface (e.g. crevasses) against the different capacity of the others spaceborne imagery.

As said before, the glaciers and snow fields normally exist in remote and inaccessible areas and the data collection on a regular basis becomes quite difficult and hazardous. The multi-spectral satellite data of Landsat MSS and TM became an important resource to mapping glaciers and associated landforms. Philip et al. [15] used visual interpretation techniques based on standard photointerpretation methods and subsequently digital image processing to retrieve glacier mapping in the 90’s decade, over the Gangotri glacier in

Figure 4.2: Surfaces of Kronebreen and Kongsvegen glaciers (Svalbard, Norway) and adjacent nunatak. a) KFA-1000 image, b) SPOT HRV-XS image, c) Landsat TM image, and d) Landsat MSS image [14].
4.1 Spatial extent and surface features

the Himalaya. They observed that the spectral reflectivity of snow is dependent on various parameters such as grain size and shape, impurity content, near surface liquid water content, depth and surface roughness and solar elevation. As spectral reflectance curves of snow in the visible and NIR wave length regions shows that in the visible region fresh snow has very high reflectance and as the snow begins to metamorphose the reflectance slightly decreases. But in the NIR region, the reflectance of aging snow decreases considerably as compared to fresh snow.

In the visible region of the EM spectrum, snow reflectance is relatively insensitive to grain size but highly sensitive to contamination by dust and soot, whereas in the NIR wavelength, the snow reflectance is sensitive to snow grain size but insensitive to contamination. As results, they found that, in the visible bands of Landsat TM, the highly reflecting surface of snow and glaciers reach saturation limits and are not useful in discriminating snow types and mapping landforms in these areas. But the TM Bands 4, 5 and 7 in the NIR and SWIR regions are found to be very useful not only in snow mapping but also in identifying various glacial landforms [15].

Later, [16] evaluate the relative accuracy of different methods for glacier mapping using Landsat TM. He considered the most used method according to the literature: manual delineation of the glacier outline, segmentation of ratio images and classification techniques. According to this, the segmentation of a ratio image of Landsat reveals the best results for glacier mapping in this test area. Other methods were too much affected by shadow, although all methods failed to discriminate debris covered ice because of its spectral similarity to surrounding terrain. Median filtering somewhat improved the accuracy, but at the expense of limiting the smallest reliably detectable glacier to 10 hectares [1].

M. Erdenetuya et al. [17] used Landsat TM and SRTM data in order to estimate glacier area obtaining the NDSI (Normalized Difference Snow Index) as input to a maximum likelihood classification approach. A temporal serie was used to calculate glacier changes. The method include the identification of spectral characteristics of glacier in each Landsat band, to apply bands combination method for glacier extraction, to apply both supervised and unsupervised classification methods, to calculated normalized difference snow index and to analyze three dimensional view of images.

Thresholding of ratio images (i.e. 4/5; 2/5; 4/2) after a time-intensive manual digitizing was found to be the most accurate method [16]. However, all such semi-automated methods failed to include debris-covered glacier areas due to spectral similarity to surrounding bedrock. Consequently, debris-covered glaciers are still mapped mostly manually, which is time intensive and therefore not suitable for studying larger areas [18].

For debris-covered glaciers or glacier parts, an additional morphometric glacier mapping approach (MGM) that focuses on curvature characteristics is capable to identify also supraglacial debris. However, results heavily depend on resolution and quality of the employed DEM as well as the specific glacier characteristics such as surface features.

Imaging radar is also effective for determining the spatial extent of terrestrial ice masses. A mosaic of SAR imagery was constructed for Antarctica in the 1990’s using Radarsat data from the Antarctic Mapping Mission (Fig. 4.3). C-band imagery can be used to map wet
4.1 Spatial extent and surface features

snow and ice-free surfaces but provides poor discrimination between glacier ice, snow, and bare rock [1].

Figure 4.3: AMM (Antarctic Mapping Mission) Radarsat mosaic of Antarctica [1].

L-band imagery distinguishes snow or ice from other surfaces, as [19] have proved. In this study, authors made an extensive review over backscattering coefficient behavior to delimit snow and glacier areas from other surfaces. Since backscattering signals from bare rock, soil and glacier are dominated by surface backscattering, the back-scattering coefficients at a given incidence angle mainly depend on dielectric properties and surface roughness. Backscattering from wet snow-covered areas is dominated by both surface and volume scattering, depending on their physical parameters and surface roughness. Generally, the dominant scattering mechanics are surface backscattering at small incidence angle and volume backscattering at large incidence angle. Since the surface of rock and soil is rougher than wet snow and glacier, the backscattering coefficients from rock or soil are greater. The roughness of glacier ice surfaces is generally smaller than rock or soil but greater than wet snow, hence the magnitude of their backscattering coefficients is between those from rock or soil and wet snow. These differences in dielectric properties and surface roughness provide an opportunity to separate glacier, wet snow and bare rock or soil regions [19].

They found that C and L band provide good capability to map wet-snow and ice-free surfaces with accuracies greater than 80%, but they poorly separate glacier ice from snow and rock. L-band HH polarization SAR UERS-I can distinguish snow or glacier from other surfaces but cannot discriminate between snow and glacier ice. Overall, TM is better than SAR’s for mapping glacier ice in alpine regions (Fig. 4.4), because the backscattering measurements from SAR have much greater fluctuations than TM reflectances. These fluctuations are caused by greater sensitivity to the surface characteristics of the glacier ice, larger
4.2 Glacier thickness

The thickness of terrestrial ice masses is easier to determine than that of sea ice and freshwater ice, because glacier ice is remarkably transparent to electromagnetic radiation in the MHz to GHz region, which is capable of penetrating kilometers of ice [1]. This provides the rationale for radio echosounding and ground-penetrating radar or impulse radar. The basis of the measurement is simply to measure the propagation time for a short radio-frequency pulse to travel to the bedrock and back. Bedrock topography can be also determined directly from knowledge of the surface topography and the ice thickness [1].

One of the newest projects to reveal ice thickness in glacier and ice sheets form remote sensing is the ESA’s CRYOSAT-2 mission for polar science. The original CryoSat satellite was lost in 2005 as a result of a launch failure. Nevertheless, the launch of the replacement satellite in April 2010 has resulted, among other studies, in the first maps of ice thickness in the Arctic and Antarctic ice sheets (Fig.4.5 and 4.6)
Figure 4.5: Antarctic ice Sheet from Cryosat-2 [20].

Figure 4.6: Arctic ice Sheet from Cryosat-2. "Satellite data since 1978 show that annual average Arctic sea-ice extent has shrunk by 2.7% per decade" (Climate Change 2007 Synthesis Report by the Intergovernmental Panel on Climate Change) [20].
4.2 Glacier thickness

Cryosat is capable to measure two types of polar ice: ice that cover land and ice floating over the ocean. As melting of floating ice has no direct effect on sea level rise, the challenge of Cryosat is to detect the melting of ice masses on land as Arctic and Antarctic, which can direct influence on the sea level rise [20].

CryoSat-2’s primary payload is the SAR/Interferometric Radar Altimeter (SIRAL), designed to meet the measurement requirements for ice-sheet elevation and sea-ice freeboard. Conventional radar altimeters send pulses at intervals long enough that the echoes are "uncorrelated"; many such echoes can be averaged to reduce noise. At the typical satellite orbital speed of 7 km/s, the interval between pulses is about 500 µs. However, the CryoSat-2 altimeter sends a burst of pulses with an interval of only about 50 µs.

The returning echoes are correlated and, by treating the whole burst together, the data processor can separate the echo into strips arranged across the track by exploiting the slight frequency shifts, caused by the Doppler effect, in the forward- and aft-looking parts of the beam. Each strip is about 250 m wide and the interval between bursts is arranged so that the satellite moves forward by 250 m each time. The strips laid down by successive bursts can therefore be superimposed on each other and averaged to reduce noise (SAR mode).

In order to measure the arrival angle, a second antenna receives the radar echo simultaneously. When the echo comes from a point not directly beneath the satellite, there is a difference in the path-length of the radar wave, which is measured. Simple geometry then provides the angle between the "baseline" joining the antennas and the echo direction. The difference in path length is tiny-up to a wavelength of the radar wave (2.2 cm) and has to be accurately determined in an overall range measurement of 720 km.

In addition to the altimeter, knowledge of the precise orientation of the baseline of the two receiving antennas is essential. CryoSat-2 measures this baseline orientation using the oldest and most accurate of references: the position of the stars in the sky. Three startrackers mounted on the antenna support structure each takes five pictures per second. Each image is analysed by the startracker’s built-in computer and compared to a catalogue of star positions. The altimeter makes a measurement of the distance between the satellite and the surface. However, this measurement cannot be converted into the more useful measure of the height of the surface until the satellite’s position is accurately known.

These days, the orbital position of altimetry satellites can be determined to within a few centimetres. To do this, CryoSat-2 carries two devices: a radio receiver and a laser retroreflector.

The Doppler Orbit and Radio Positioning Integration by Satellite (DORIS) radio receiver detects and measures the Doppler shift on signals broadcast from a network of more than 50 radio beacons around the world. Although the full accuracy of this system is obtained only after ground processing, DORIS provides a realtime estimate on board, good to about half a metre. The DORIS system has been operating for more than a decade, and is used on many satellites, such as Envisat. The small laser retroreflector is attached to the underside of CryoSat-2. This little device has seven optical corner cubes, which reflect light in exactly the direction it came from. A global network of laser tracking stations fires short laser pulses at CryoSat-2 and times the interval before the reflected pulse arrives back.
These stations are relatively few but, because their positions are very accurately known, from their routine work of tracking geodetic satellites, they provide a set of independent reference measurements of CryoSat-2’s position.

4.3 Interferometric applications for Glacier monitoring

Interferometric SAR is an alternative to conventional stereographic techniques for generating high resolution topographic maps, and can also be used for velocity and change detection mapping [21].

The use of spaceborne SAR’s as interferometers (interferometric SAR=InSAR or IF-SAR) became popular only recently, although the basic principle dates back to the early 1970’s.

Interferometry SAR exploits the phase differences of at least two complex-valued SAR images acquired from different orbit positions and/or at different times. It uses the measured differences in the phase of the return signal between two satellite passes to detect slight changes on the Earth’s surface. The combination of two radar measurements of the same point on the ground, taken at the same time, but from slightly different angles, produces stereo images. Using the cosine rule from trigonometry to calculate the distance between the radar and the Earth’s surface, these measurements can produce very accurate height maps, or maps of height changes [22].

The differential interferometry (DinSAR), an alternative technique, represents a unique method for detection and mapping of surface displacements over large temporal and spatial scales with precision in the cm and even mm range. This is of great importance for earthquake and volcanic research, glaciology and ice sheet monitoring, studying tectonic processes, monitoring land subsidence due to mining, gas, water, and oil withdrawal, etc [22].

Next sections will discuss main applications related to glacier monitoring where radar interferometry already demonstrated its capability to provide spatial change maps.
4.3 INTERFEROMETRIC APPLICATIONS FOR GLACIER MONITORING

4.3.1 Glacier velocity mapping

As said before, topographic mapping with InSAR relies on acquiring data from two different look angles and assumes that the scene imaged did not move between data acquisitions. If the look angles of multiple data acquisitions are identical, equivalent to having a zero baseline, there is no sensitivity to topography, and the interferogram can be used to extract information regarding scene dynamics [21].

The possibility of using SAR interferometry to determine glacier velocity was discovered in the 1990’s after the launch of ERS-1 [3]. After that, many applications in the framework of interferometric principles have been developed for glaciers monitoring. In the next lines, some approximations to detect glacier movements and velocity will be exposed.

A SAR system measures the amplitude of the signal, influenced by the target properties, and the time delay of the signal, which gives the distance between the sensor and the target (this distance is called the range). The SAR records the phase of the reflected wave, and so that two images taken from an orbit separated by a small distance will each have a different distance from the sensor to a same target. Subtracting the phases of both images results in an interferogram, whose phase contains the difference in range. The interferogram has its phase value displayed in every pixel; thus phase variations (i.e., fringes) from pixel to pixel are visible [3] (Fig. 4.7).

Assuming no movement of a target, the phase in an interferogram (see Appendix A) is influenced by geometric effects and the topography. For a moving target, the phase is additionally influenced by the translation of the target. The phase due to topography has to be extracted to get the velocity. The synthetic interferogram is then subtracted from the real interferogram, resulting in an interferogram that contains only the phase term due to translation [3].

One of the main advantages of interferometry is to give velocity values at any point of the glacier, while other techniques like "feature tracking\(^1\)", depends on visible surface features and may not be possible in homogeneous areas without distinct surface structures, such as undulations or crevasses.

\(^1\)Morphological pattern analysis in the visible range on satellite images. This is the classic method to calculate glacier velocity, through sequential satellite imagery by observing the movement of surface features, such as crevasses.
4.3 Interferometric applications for Glacier monitoring

Figure 4.7: SPOT satellite image (left) and SAR interferogram of Kronebreen glacier, Svalbard, Arctic Glacial Ocean (right). Parallel colored bands (fringes) show movement of the glacier, moving perpendicular to these fringes. No fringes are seen for the terrain, since this effect has been removed. Fringes on the ocean are due to the moving sea ice [3].

D. Floriciou et al [23] investigated the capabilities of TerraSAR-X data for feature tracking by amplitude correlation over Moreno, Upsala and Ameghino glaciers of Southern Patagonia Ice Field in Argentina, which are calving into the Argentino lake. Correlation techniques with 11-days repeat-pass, between December 2007 and January 2008, amplitude TerraSAR-X images is applied for ice motion studies. They used an incoherent amplitude correlation method, based in geocoded products (with 30 m. SRTM DEM). As correlation measurements are unambiguous, lower values than 0.4 m/11 days = 0.036 m/day, are considered as orbital errors. Then displacement between two successive images (11 days) is on the order of 55 meters, i.e. about 50 pixels in the geocoded product.

The amplitude correlation method delivers a 2D displacement vector in range and azimuth. Since they perform the correlation on geocoded products, the displacement vector is derived directly in ground geometry. Before that, a low pass filtered resolution image is then subtracted to enhance the image contrast which stabilized the correlation. Finally, when coherence and displacement vectors, a coloured overlay image is computed by adding the colored velocity map on the grey scale radar intensity map. The overlay is weighted with the coherence of the estimate.
Figure 4.8 exposes the processing flow chart of the SAR amplitude correlation method and the TerraSAR-X derived map for glacier velocity estimation of Upsala glacier. The frontal velocities reach maximum values of 5.6 m/day which are larger than the maximum value of 4.9 m/day measured in 1993 at stakes in similar location [23]. The acceleration is very likely caused by effects of reduced friction at the glacier bed. Due to decreasing ice thickness the grounded terminus approaches flotation thickness near the front. This coincides with major frontal retreat during the last several years [23].

N. Gourmelen et al. [24] combined conventional InSAR and Multiple Aperture InSAR (MAI) to determine the ice surface velocity on the Langjokull and Hofsjokull ice caps in Iceland in 1994. MAI is a new technique applicable to SAR data from all available SAR instruments. MAI extends the sensitivity of InSAR to along-track displacement, thus allowing a complete recovery of the three-dimensional (3D) displacement field, using a single pair of SAR data and parallel flow assumption or two pairs of SAR images. Using ERS-1 data, they have produced two conventional interferograms and two MAI interferograms, as well as two speckle tracking offset maps for comparison purposes. To reduce topographic effects Digital Elevation Model have been used.

Glacier velocity with feature tracking can be detected both by SAR interferometry and optical data. In the second case, glacier velocity can be determined simply by visual obser-
4.3 INTERFEROMETRIC APPLICATIONS FOR GLACIER MONITORING

vation, using sequential satellite imagery by observing the movement of surface features in time, such as crevasses, co-registering sequential images. For example, one in the red channel, one in the green channel; features that have moved will appear as tones of green or red. Then, moving of one image over the other until a moved feature matches again results in a yellow tone for this feature, and the moved distance can be determined [3].

Scambos et al [25] developed an automated method which was successfully applied in many studies in Antarctica, called IMCORR [26]. The method takes two images and a series of input parameters and attempts to match small sub scenes (called chips) from the two images. The program uses a fast fourier transform (based version of a normalized cross-covariance method).

The most common use of this type of algorithm in image processing is to accurately locate tie-point pairs in two images to coregister them. However, if the images are already coregistered by other means, the algorithm may be used to find the displacements of moving features, provided that the features show little change in their appearance, and that the motion is strictly translational. IMCORR takes as input the image names and sizes, parameters determining search chip size, reference chip size, grid spacing, and output filename. Further, preset offsets of search chip centers may be specified, and subareas of the full image files may be used to restrict the area over which IMCORR attempts to find displacements. At each of the gridpoints IMCORR calculates a correlation index for every location at which the reference chip will entirely fit within the search chip.

IMCORR takes the correlation values in the vicinity of the best integer-pixel match and interpolates a peak correlation location to sub-pixel precision. The program returns a file containing the locations of the grid centers for the reference chips, the displacements required to best match the chip pairs (or indicates that none could be found), and several quality control parameters that may be used to evaluate the validity of the match. We use this program to measure glacier velocities; however, the same program may be useful for other applications.

Scherler et al [27] applied a novel technique for precise orthorectification, co-registration, and sub-pixel correlation of Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) satellite imagery to derive surface velocities of Himalayan glaciers. Their approach includes correction of offsets in the displacement maps due to attitude effects and due to elevation errors in the DEM. The methodological principles are applicable to a wide variety of optical satellite imagery and are demonstrated using ASTER images. They have studied the glaciers in two Himalayan regions: Khumbu in Nepal and Garhwal in India, where the glacier shrinking is observed. First, they demonstrate the methodological principles, including quality assessment, on the relatively slow Khumbu glacier at Mount Everest. Second, they investigate and model displacement errors induced by systematic elevation errors in the SRTM-based DEM, at the Gangotri glacier group in Garhwal. In a further step, the recent velocity history of Gangotri glacier, situated in the headwaters of the Ganges, is analyzed to demonstrate the capabilities and the limits of the method to monitor glacier dynamics.

The first group of tasks comprises orthorectification, co-registration, and correlation of
the satellite imagery, followed by post-processing of the correlation results using COSI-Corr. COSI-Corr is a IDL-based module for the remote-sensing platform ENVI by RSI. The application allows processing satellite imagery from the SPOT, ASTER, and Quickbird sensors. The second group of tasks is related to data filtering and assessing the quality of the results. In case of more than one correlation, i.e., more than two ortho-images, further steps may involve the comparison and the combination of the acquired data.

4.3.2 Glacier mass balance

The most obvious indicator of changes in a glacier is the advance or retreat of front position, which is caused by changes in glacier mass balance. The response time to such changes, however, is highly variable, depending of both direct climate forcing and individual physical conditions such as glacier size, topography and ice temperature [3].

Glacier mass balance study is concerned with changes in glacier mass through time, and specially the changes from year to year. Glacier mass balance forms the vital link between the changing atmospheric dynamics and glacier dynamic and hydrology. The most obvious present-day connection between glacier mass balance and global climate change is the rise of global sea level that will occur if glacier melting increases in response to global warming [28].

Mass balance (or mass budget) is the change in the mass of a glacier or ice body, or part thereof, over a stated span of time. The span of time is often a year or a season. A seasonal mass balance is nearly always either a winter balance or a summer balance, although other kinds of season are appropriate in some climates, such as those of the tropics. The definition of "year" depends on the method adopted for measurement of the balance [29].

The (cumulative) mass balance, \( b \), is the sum of accumulation, \( c \), and ablation, \( a \) (the ablation is defined here as negative) (Eq. 4.1). The symbol, \( b \) (for point balances) and \( B \) (for glacierwide balances) has traditionally been used in studies of surface mass balance of valley glaciers (mass balance is often treated as a rate, \( b \) or \( B \) dot).

\[
b = c + a = \int_{t_1}^{t} (c + a)\,dt \tag{4.1}
\]

Accumulation can be defined as the mass gained by the operation of all processes that add to the mass of the glacier, expressed as a positive number, such as snow fall (usually the most important), hoar (a layer of ice crystals, usually cup-shaped and faceted (see Appendix B)), freezing rain, gain of windborne blowing snow and drifting snow (can be important for the survival of, for example, small cirque glaciers), and avalanching. On the contrary, ablation is the process that reduce the mass of a glacier, hence, expressed as a negative number. The main components of ablation are melting (usually the most important on land-based glaciers), calving (icebergs discharge into seas or lakes, approximately 50 and 90 of ablation occurs via calving in Greenland and Antarctica), loss of windborne blowing snow and drifting snow, avalanching and sublimation (for example, at high alti-
4.3 Interferometric Applications for Glacier Monitoring

tudes in low latitudes, tropical glaciers, in dry climates, and on blue-ice zones in Antarctica; is a function of vapour pressure) [29].

**Annual mass balance** \( (b_a) \) is the mass balance at the end of a **balance year**. It can also be described as the sum of the winter balance \( (b_w) \) and summer balance \( (b_s) \), which is negative (Eq. 4.2).

\[
b_a = b_w + b_s \tag{4.2}
\]

Mass balance can be calculated at a point (indicated by small letters, \( b_a \)) or volume for the whole glacier (indicated by capital letters, \( B_a \)). The annual balance \(^2\) \( B_a \) of the entire glacier with area \( A \) is given by Eq. 4.4:

\[
B_a = \int b_a dA \tag{4.3}
\]

Accumulation and ablation are usually seasonally governed, and so the mass balance undergoes an annual cycle of growth (positive mass balance) and diminishment (negative mass balance). At mid-latitudes, there are distinctly different accumulation and ablation seasons, i.e. it snows mainly during the winter and melts during the summer. This is different from e.g. tropical regions, where ice melting and accumulation occur at the same time. Annual mass balance is usually expressed as illustrate Figure 4.1.

![Figure 4.9: Winter (white), summer (grey), and net mass balance (black) for Storglaciaren (Sweden) 1946-2002. Black lines show the running 5-year average [29].](image)

Variations in area and volume are interesting because they are ways easy to get a good idea of the evolution of the mass of a glacier, which is the ultimate variable to understand.

\(^2\)Mass balance terms are stated as **water equivalent (w.e.)**, so that comparisons can be made between different glaciers and different years. Water equivalent represents the volume of water that would be obtained from melting the snow or ice. The value in 3 meters water equivalent is obtained through dividing the volume by the area, i.e. the mass balance value states how much the glacier has become thicker or thinner (in water depth) if the mass addition or loss is distributed over the whole glacier surface [29].
its dynamics. Thus the glacier mass balance is essential to address issues on the status and future of a glacier.

There are several ways to estimate this mass balance, whether considering the variation of volume and density, what is known as the geodetic method, or quantifying processes accumulation and ablation on the surface, which constitutes the classical glaciological method. Traditionally it has been determined by glaciological method, using labour-intensive point measurements of ablation and accumulation distributed over the glacier. Although ground-based, in-situ techniques exist for measuring glacier mass balance they tend to be labour-intensive, expensive and, usually, provide very limited spatial coverage. There are several characteristics of the surface of a glacier that can be derived from remote sensing data and which, in some way, may be useful for determining mass balance.

Recession of a glacier snout is often cited as evidence for a negative mass balance but should, in general, be treated with caution as changes in terminus position may or may not reflect changes in mass balance further up-glacier and, in particular, in the accumulation area. Thus, determining a volume or mass change, integrated over the whole glacier, can be problematic, especially in the case of calving glaciers [30].

Rott et al [31] developed a spatially detailed glacier mass balance model that calculates daily accumulation and ablation in dependence of surface elevation, using satellite data from ASAR and MODIS sensors. They design a glacier mass balance model called GMB-RS based on software developments and experience in snowmelt runoff modeling and forecasting at ENVEO for Alpine basins. The model concept combines the temperature index approach (classically used as input for mass balance analysis) with satellite observations.

Figure 4.10 illustrates the basic concept of GMB-RS. Satellite data are used for model set-up, as well as for the continuous model runs calculating surface mass balance in daily steps. Glacier boundaries are mapped in high resolution optical satellite images. Glacier surface topography can be retrieved by means of SAR interferometry, or obtained from other sources. The Shuttle Radar Topography Mission (SRTM) was used for glacier model set-up on the Patagonia Icefield, where no other reliable topographic data are available. interferometry of the ERS-1/ERS-2 tandem mission. For estimating mass balance of calving glaciers, data on ice velocity at the front and ice thickness are needed to calculate the ice export due to calving, using also interferometry.
Figure 4.10: Basic concept of GMB-RS, based on meteorological data and EO satellite data [31].

For calculating snow and ice melt with GMB-RS, a temperature index is used which is a simple parameter for the energy available for melt. The amount of snow or ice melted during a time period is set equal to the sum of positive air temperatures \( T^+ \) (°C) over this period, times a factor of proportionality, the positive degree-day factor, DDF. The factor is different for snow and ice, and varies also with the saturation of the snowpack and with the surface albedo. DDF can be calibrated accurately by means of ablation measurements in the field. However, as the use of mass balance models is primarily aimed at glaciers without field observations, DDF has to be estimated from observations on other glaciers in the region or from published data.

The model calculates the net balance \( B_n \) (m³) (Equation 4.4) in dependence of elevation of the glacier surface. For elevation zone \( i \) the net balance during time step \( t \) is:

\[
B_{n,i}(t) = C_{sn,i}DDF_{sn,i}(t)T_i^+(t)A_{sn,i}(t) + C_{ice,i}DDF_{ice,i}(t)T_i^+(t)A_{ice,i}(t) + f_p(T)C_pP_i(t)
\]  

(4.4)

The index \( sn \) refers to the snow surfaces in the elevation zone, and ice to the ice surfaces. \( T^+ \) is the daily mean air temperature above a threshold close to 0°C (degree days) at the hypsometric mean height of the zone. \( A \) is the area of snow or ice, respectively, and \( C_{sn} \) and \( C_{ice} \) are correction factors accounting for losses (e.g., evaporation). \( P \) is the precipitation, and the factor \( f_p \) decides if \( P \) falls as rain or snow. For calculating snowfall we use a transition with varying percentage of solid and liquid precipitation for mean daily temperatures between -1°C and +2°C. The coefficient \( C_p \) accounts for the percentage of precipitation that is stored in the snow or firm.
Although the model has been tested in other areas as well, an example is exposed subsequently for glaciers of Patagonia Icefields: the two adjoining eastern outlet glaciers, Moreno glacier (MG) and Ameghino glacier (AG). MG covers 254 km$^2$ in area and flows over a length of about 30 km from the continental divide, with Cerro Pietrobelli (2950 m a.s.l.), down to Lago Argentino at 185 m.

Glaciological field measurements were carried out on the glacier from 1995 to 2003. Whereas ice ablation was measured directly by means of ablation stakes, total net accumulation was determined from the mass transport through a transverse profile. The calving flux was determined using satellite-derived surface velocities and lake depth in front of the glacier. Moreno glacier is close to steady state, as the damming events in 2003/04 and 2005/06 have shown when after minor frontal advance the glacier dammed the southern arm of Lago Argentino for several months.

Ameghino glacier, on the other hand, shows significant retreat since about 40 years. It calves into a proglacial lake that started to form in the late 1960’s. On AG only few ablation stake measurements are available over a one year period, so that the mass balance estimate has to rely on model calculations. We carried out these calculations for Ameghino main glacier, covering an area of 53 km$^2$.

Figures 4.11 and 4.12 show the surface mass balance for both glaciers for a typical year of the observation period.

![Figure 4.11](image1)

Figure 4.11: Modeled specific and absolute ($10^9$ kg/100 m. elev.) annual surface mass balance for Moreno glacier [31].

![Figure 4.12](image2)

Figure 4.12: Modeled specific and absolute ($10^9$ kg/100 m. elev.) annual surface mass balance for Ameghino glacier [31].
Disasters associated to the glacial and periglacial environment can cause thousands of casualties in one event. The present shift of cryospheric hazard zones due to atmospheric warming, process interactions and chain reactions, and the potential far reach of glacier disasters make it necessary to apply modern remote sensing techniques for the assessment of glacier and permafrost hazards in high mountains.

Application of earth-observation techniques is important because, typically, related hazard source areas are situated in remote high-mountain regions, often difficult to access physically for topographic, political and/or security reasons; the remote location of most glacial hazard sources, the potential process interactions and chain reactions, and the far reach of some of the high-mountain hazards require remote sensing sensors capable to cover large areas at once; and due to the current rapid change of high-mountain environments, hazard assessments shall be undertaken routinely and regularly, combined with continuous monitoring.

In this chapter an overview of different hazard types and how remote sensing methods can be applied for their assessment is provided, according to [32].

5.1 Glacier-related floods

Glacier floods represent the glacial risk with the highest potential for disaster and damages. They occur in most glacierized mountains of the world and are triggered by the outburst of water reservoirs underneath and at the margins of glaciers. Most reservoir types develop slowly and can be identified at the surface, a precondition that favours the application of remote sensing techniques for monitoring glacial and periglacial lakes. Floods from ice-
dammed lakes and proglacial moraine-dammed lakes, in particular, represent a recurring and severe danger. Different outburst mechanisms are involved in glacier floods:

**Breaching of moraine dams:** Trigger for such process are usually enhanced runoff into the glacial lake, impact waves (e.g. from ice, rock or snow avalanches), temporary damming/jamming at the lake outlet, or progressive flow and erosion in the moraine (piping). Moraine-dammed lakes are usually detectable by remote sensing, in particular optical techniques. Time series of images are particularly useful for assessing lake dynamics and estimating future development. The assessment of moraine dam characteristics requires high resolution and precision techniques (dam geometry, deformation, settlement, surface material, etc.). Monitoring of associated glacier characteristics (geometry, surface type), changes and kinematics (thickness changes, velocity) may help assessing the evolution of proglacial lakes.

**Failure or overtopping of ice-dams:** For permanent ice dams, the disasters often repeat because the ice dam may recover after an outburst. Related outbursts stem from ice-marginal lakes and from temporary ice dams from ice avalanches or glacier surges. The detection of ice-dammed lakes is depending on the temporal resolution and timing of the remote sensing system applied, the detection of ice dams is depending on the spatial resolution and spectral characteristics. Time series are particularly useful. The monitoring of thickness changes and kinematics of longlasting ice dams may support the assessment, e.g. of the floatation level.

**Glacier outbursts:** are catastrophic water discharges from the subglacial drainage system. Glacier outbursts are particularly difficult or impossible to assess due to their sub-surface character. An exception from this problem is in some cases the detection of strong sub-glacial volcanic activity.

**Breaching of thermokarst and supra-glacial lakes:** Thermokarst and supra-glacial lakes can develop on ice-rich permafrost or stagnant glacier ice. Thermal convection leads often to progressive lake growth. Outburst causes are similar to breaching of moraine and ice dams, and, in addition, progressive melt of the ice or permafrost dam. Detection of related lakes usually requires high image resolution due to the small lake size. Time series can be particularly useful. The disposition of lake development is partially detectable through remote sensing of surface characteristics and kinematics.

**Displacement waves:** Displacement-waves impact on people, natural and artificial lake dams, and other installations. Those displacement waves have been the trigger for a number of lake outburst events. They originate from the impact of snow, ice, rock-avalanches, landslides, debris flows, etc. into the lake, or from floatation of icebergs. Their assessment requires integrated remote sensing and modelling approaches of source processes.
5.2 Glacier length and volume changes

Advancing and retreating glaciers can pose a direct risk to mountain infrastructures. From a global point of view, the prediction of glacier length variations is complicated by the fact that glaciers can vary in a continuous (stable) or unstable way (i.e., glacier surges).

**Glacier surges:** are a temporary instability of large glacier parts with ice velocity increased by an order of magnitude (or more). Usually, glacier surges are accompanied by drastic glacier advance. Surging glaciers are able to rapidly destroy installations or induce other hazards, such as ice- and rock-falls. They can temporarily dam lakes, which, when these dams fail, produce some of the largest known outburst floods. Often, glacier surges are accompanied by enhanced englacial water storage, which is possibly released at the surge end in a catastrophic way. Surges can be tracked by high-frequency remote sensing. Former glacier surges, and thus surge-type glaciers, can often be recognised from deformed moraines. Geometry changes, if involved in the surge disposition and build-up, may be detected as glacier thickness changes.

**Stable glacier advance and retreat:** Advancing glaciers may inundate land, override installations, dam rivers and form lakes, cause ice break-offs, etc.. Glacier retreat forms usually no direct hazard but is able to trigger a number of secondary hazards such as various slope instabilities and ice avalanches. Causes for such stable glacier length changes are changes in mass balance and/or in ice dynamics. Stable glacier length changes can usually be monitored by remote sensing: glacier area changes from repeat imagery and glacier mass changes from repeat digital terrain models (DTM). Forecast of such length changes is best done by a combination of remote sensing, glaciological field work and modeling.

**Changes in glacier runoff and seasonality:** Glacier mass loss leads to reduction of water resources as stored in glaciers and to changes in dry-season river flows. The short-term perspective is increasing discharge due to enhanced melt; the long-term perspective is decreasing discharge when the glaciers in a basin become substantially smaller or disappear. These processes have consequences for drinking water supply, irrigation, hydropower production, industrial water use, fishery, water quality, etc. Changes in glacier runoff are best investigated through a combination of remote sensing, meteorology, and combined glaciological and hydrological modeling.

5.3 Glacial and paraglacial mass movements

Compared to the distances covered by glacier floods, ice and rock avalanches often affect much smaller areas. Corresponding disasters are generally restricted to densely populated high-mountain regions. However, in combination with other hazards, ice and rock avalanches have the potential for far-reaching disasters. In zones with high seismic activ-
ity and geothermal heat flow, the risk of major ice break-offs is greatly increased, as was
demonstrated dramatically by one of the most destructive glacier catastrophes, the Huas-
caran disaster in 1970, with a loss of over 18,000 lives. Also, the extraordinary 20 Septem-
ber 2002 rock/ice avalanche at Kolkä/Karmadon (Caucasus), a combination of rock and ice
destabilisation killing over 100 people, drastically underlines the devastating potential of
ice/rock avalanches. A widespread risk in high mountains is related to accumulations of
loose sediments on steep slopes, which represent potential sources of debris flows. Such
debris accumulations can occur in the form of moraines, moraine dams, or steep valley
flanks uncovered by retreating glaciers.

**Ice fall and ice avalanches:** Steep glaciers (or glacier parts) are the usual source of
ice break-offs and subsequent ice avalanches. In rare cases the detachment of complete
glaciers seems possible. Ice avalanches are particularly dangerous in winter with reduced
basal friction, extended run-out, and mass gain from snow. Glacier parts can also fail due
to a failure of the underlaying rock. Ice avalanches can be triggered by earthquakes. Ice
avalanches itself can trigger lake outbursts, dam rivers, or transform into mud/debris flows.
Detection of steep glaciers is possible through combination of spectral data with a DTM.
High-resolution, -precision, and -frequency remote sensing (e.g. terrestrial close range
techniques) enables sometimes the monitoring of mass changes and kinematics related to
entire steep glaciers or unstable sections.

**Rock fall and rock avalanches:** Glacier retreat uncovers and debuttresses rock flanks.
The related change in thermal, hydrologic, hydraulic and mechanic conditions can lead
to rock fall and rock avalanches (fast mass movement). Rock avalanches can carry parts
of overlaying glaciers. Rock avalanches can be of increased magnitude in glacial envi-
nonments (extended runout on glaciers, or when combined with ice; mass gain from ice;
entrainment and liquifaction of glacier parts through impact and transport; detachment of
glaciers overlaying the rock mass breaking off). Rock avalanches can be triggered by earth-
quakes. Mapping of rock faces and some boundary conditions (e.g. glacier retreat) are
possible through remote sensing.

**Landslides and rock slides:** Among other causes, glacier retreat or slope undercut-
ting by floods uncovers and debuttresses rock and debris flanks. The related change in
hydrologic, hydraulic and mechanic conditions can lead to mass movements (slow mass
movement). These can create secondary hazards such as river dams. Landslide surface
characteristics, geometry and kinematics can be monitored by repeat high-resolution and
-precision remote sensing.

**Destabilisation of unconsolidated glacial deposits:** Glacier retreat leaves unprotected
and unconsolidated moraine material that is prone to enhanced erosion and debris flows.
Related zones can be detected through remote sensing combined with DTMs.
Debris flows from glacier floods: Glacier and permafrost floods are often accompanied by debris flows when erodible material is available in steep parts of the flood path.

Interaction between volcanic activity and glaciers: Interactions between volcanic activity and glaciers are potentially among the most devastating disasters with glacier involvement. Enhanced geothermal activity, geometric and mechanic changes, deposition of hot eruptive materials, or albedo change by volcanic ash can lead to drastic melt of ice or ice break-off on ice-clad volcanoes and to volcanic landslides or lahars. Ash layers thicker than some mm or cm insulate the underlying ice. Ice cover on volcanoes and its changes (and partially also volcanic activity) can be monitored by remote sensing.
As has been discussed in literature, satellite remote sensing is valuable and complementary technique for studying and monitoring snow and glacier ice, having the capability to conduct research over large and otherwise inaccessible areas.

Among others, the properties that can be measured are surface extent, thickness, glacier velocity, mass balance and applications such as to emergency event related to glaciers. These measurements are used in global, regional and local scales, thorough remote sensing and ground data for precise interpretation.

This report has done a review of the study of this main applications, making a conceptual introduction of each process and describing the existent approaches from remote sensing tools. The methods presented have the limitations of present satellite sensors, such as new and more robust algorithms needs to be developed in the future for more precise and accurate results.

Finally, the most important conclusion is related to the fact that remote sensing became a very useful tool for this topic, even more if climate behavior continues being the key of the glaciers retreat problem.
Interferometry principles

Note: This Appendix, made according to [33, 34, 21], has the purpose of reviewing interferometric main concepts in order to understand its principle for glacier monitoring.

One of the more interesting applications of synthetic aperture radar imagery to emerge in the past two decades has been topographic mapping and change detection using interferometry. Because the phase angle of the backscattered signal for a given pixel is available, and phase is easily measured, it is possible to compare the phase differences of two different images of the same region and, from that comparison, find the relative locations of pixels in three dimensions: latitude, longitude and altitude, or their equivalents.

A SAR works by illuminating the Earth with a beam of coherent microwave radiation, retaining both amplitude and phase information in the radar echo during data acquisition and subsequent processing. This radiation can be described by three properties:

- **Wavelength**: the distance between peaks on the wave,
- **Amplitude**: the displacement of the wave at the peak, and
- **Phase**: describes the shift of the wave from some other wave (measured in angular units, like degrees or radians).

Synthetic Aperture Radar (SAR) interferometry exploits this coherence, using the phase measurements to infer differential range and range change in two or more complex-valued SAR images of the same surface, thereby deriving more information about an object than is obtainable with one single image. The resulting difference of phases is a new kind of image,
A INTERFEROMETRY PRINCIPLES

called an interferogram, which is a pattern of fringes containing all of the information on relative geometry.

For a second SAR image to provide additional information, it must be acquired from a different sensor position or at a different time. The difference between the acquisitions of the first and second images determines the type of interferometer that results. Some of the most common forms are:

- Across-track; used primarily for topographical information, this type utilises a difference in across-track position, or look angle (range direction),

- Along-track; used primarily for ocean currents information and moving object detection, this type utilises a difference in the along-track position (azimuth direction), which can be achieved by a small difference in acquisition time, on the order of microseconds to seconds,

- Differential; this method utilises a difference in time, on the order of days to years, and is used primarily to observe glacier (ice field) or lava flows, if the time difference is within days. If the time difference is measured in days to years, it can be a very useful method of observing subsidence, seismic events, volcanic activity, or crustal displacement.

Across-track Interferometry (InSAR): The best known application of SAR Interferometry is the reconstruction of the Earth’s topography by using different look angles to compare the same object. This is what is referred to as across-track interferometry. Across-track is also known as the range direction, defined as the dimension of an image perpendicular to the line of flight of the radar.

Consider two radar antennas, A1 and A2, simultaneously viewing the same surface and separated by a baseline vector \( \mathbf{B} \) with length \( B \) and angle \( \alpha \) with respect to horizontal. A1 and A2 may also represent a single antenna viewing the same surface on two separate passes (Figure A.1). A1 is located at height \( h \) above some reference surface. The distance between A1 and the point on the ground being imaged is the slant range \( \rho \), while \( \rho + \delta \rho \) is the distance between A2 and the same point.
The phase of pixel value in a complex SAR image depends on the scattering mechanism in the resolution cell, and the distance from the antenna to the point. If the scattering mechanism in the two images is similar, then the phase difference between the two complex SAR images is proportional to the difference in slant range from the two antenna to the point.

The similarity in scattering mechanism in the two images is indicated by the correlation coefficient of the image, which is also called coherence\(^1\) in SAR interferometry literature. Any dissimilarity of the scattering mechanism between the two images, indicated by a low coherence, results in phase noise. A certain loss of coherence results from the different look angles from two antenna to the point, and from receiver noise. Coherence loss can also result from changes in the surface between acquisitions, in the case of repeat-pass interferometry.

The slant range difference is proportional to the full phase or absolute phase of the complex-valued interferogram. However, the measured phase values of the interferogram can only take values between 0 and 2\(\pi\). That is, the phase is ‘wrapped’. Thus, in order to compute the slant range difference which is needed to compute topography, the 2\(\pi\) ambiguity inherent in the phase measurements must be solved, using techniques of ‘phase-unwrapping’.

---

\(^1\)Coherence is the fixed relationship between waves in a beam of electromagnetic (EM) radiation. Two wave trains of EM radiation are coherent when they are in phase. That is, they vibrate in unison. In terms of the application to things like radar, the term coherence is also used to describe systems that preserve the phase of the received signal. In an interferogram, coherence is a measure of correlation. It ranges from 0.0, where there is no useful information in the interferogram; to 1.0, where there is no noise in the interferogram (a perfect interferogram). Both extremes are rarely seen. Coherence is affected by local slope (steep slopes lead to low coherence), properties of the surface being imaged (vegetated or moving surfaces have low coherence), time lag between the passes in an interferogram (long lags lead to low coherence), the baseline (large baselines lead to low coherence), technical details of the generation of the interferogram (poor co-registration or resampling leads to low coherence). Coherence can serve as a measure of the quality of an interferogram; tell you more about the surface type (vegetated vs. rock); or tell you when a tiny, otherwise invisible change has occurred in the image, and it is only visible in the phase image of an interferogram \([33]\).
Once the absolute phase of each pixel of the interferogram is known, the geometry of the figure can be used to compute the topography $z(y)$, if the baseline vector $\mathbf{B}$ in Fig. A.1 is known. The result is a Digital Elevation Model (DEM) of the observed area.
Cryosphere: Word derived from the Greek krios meaning cold to denominate the Earth’s snowy and icy regions.

Depth hoar: Ice crystals (usually cup-shaped, faceted crystals) of low strength formed by sublimation within dry snow beneath the snow surface; a type of hoarfrost. Associated with very fast crystal growth under large temperature gradients. This is one way in which firn formation may begin.

Equilibrium Line Altitude (ELA): Is the equilibrium line separating the ablation area of a glacier from the accumulation area. It marks the area where accumulation is balanced by ablation. Is usually defined by direct measurements on the field, at the end of a hydrological field (at the end of a summer). ELA is a product of climate, or the weather conditions of a particular year, and changes annually in close agreement with the annual or net mass balance.

Glacier: a mass of surface-ice on land which flows downhill under gravity and is constrained by internal stress and friction at the base and sides. In general, a glacier is formed and maintained by accumulation of snow at high altitudes, balanced by melting at low altitudes or discharge into lakes or the sea.

Iceberg: A mass of ice formed from fresh water floating in the ocean. They are breakaway fragments from ice sheets or glaciers. Due to the buoyancy, only 10% of the mass appears above the surface.

Ice cap: dome-shaped ice mass with radial flow, usually covering the underlying topography.
**Ice sheet:** a mass of land ice of continental size, and thick enough to cover the underlying bedrock topography. Its shape is mainly determined by the dynamics of its outward flow. There are only two continental ice sheets in the modern world, on Greenland and Antarctica; during glacial periods there were others.

**Ice shelf:** a thick, floating slab of freshwater ice extending from the coast, nourished by land ice. Nearly all ice shelves are located in Antarctica.

**Interferometric Synthetic Aperture Radar (InSAR):** SAR interferometry is a technique involving phase measurements from successive satellite SAR images to infer differential range and range changes for the purpose of detecting very subtle changes on, or of, the Earth’s surface with unprecedented scale, accuracy and reliability. SAR interferometry has been demonstrated successfully in a number of applications, including topographic mapping, measurement of terrain displacement as a result of earthquakes, and measurement of flow rates of glaciers or large ice sheets. The term InSAR, is most commonly associated with repeat-pass interferometry. In contrast, DinSAR is used to described differential interferometry.

**Firn:** Nevé on a glacier that survives the year’s ablation season. With time much of the firn is transformed into glacial ice.

**Firn Limit:** The lower boundary of the zone of accumulation on a glacier where snow accumulates on an annual basis. Also called the Firn Line.

**Permafrost:** Zone of permanently frozen water found in high latitude soils and sediments. Five types of permafrost have been recognized: continuous permafrost, discontinuous permafrost, sporadic permafrost, alpine permafrost, and subsea permafrost (form of permafrost that exists beneath the sea in ocean sediments).
Bibliography


[20] ESA (European Space Agency), *CRYOSAT ESA’s Ice Mission*.


