Table S5 Brief summary of studies of the January Thaw

Study	Geographical coverage	Period of record	Analyzed variables	Conclusions
Esten and Mason (1910)	Storrs, CT	1888–1909	Record max and min; max, min and avg. temps	Singularity
Marvin (1919)	Continental US	1778–1865	Avg. weekly temps	No strong singularity
Nunn (1927)	Northeast US	1873–1925	Avg. temps	Singularity
Slocum (1941)	Northeast US	1871–1939	Avg. temps	No conclusion
Wahl (1952)	Northeast US	1873–1952	Avg. temps, sea-level pressures	Singularity
Wahl (1953)	Boston, MA	1873–1952	Avg. temps	Singularity
Brier (1954)	N. hemisphere	1899–1939	Sea level pressure	No conclusion
Lautzenheiser (1957)	Boston, MA	1911–1950	Avg. temps	No strong singularity
Dickson (1959)	Nashville, TN	1871–1950	Avg. temps, tornadoes	Singularity
Bingham (1961)	Northeast US	1896–1956	Avg. weekly temps	No strong singularity
Duquet (1963)	Northeast US	1872–1961	Weekly avg. temps	Singularity
Newman (1965)	Boston, MA	1872–1964	Max and min temps	No strong singularity
Frederick (1966)	US and SW Canada	1897–1956	Avg., max, and min temps	Singularity
Hayden (1976)	East Coast US	1954–1970	Means of surf heights	Singularity
Logan (1982)	Portland, Maine	1965–1979	Avg. temps	No conclusion
Lanzante and Hernack (1982)	New Brunswick, NJ	1858–1981	Max temps	Singularity
Lanzante (1983)	N. America, Atlantic and Pacific Oceans	1947–1976	700-mb heights	Singularity
Kalnicky (1987)	N. hemisphere	1899–1969	Sea-level pressure	No conclusion
Guttman and Plantico (1987, 1989)	Eastern US	1951–1980	Max and min temps	Singularity
Guttman (1991)	Central Park, NY	1876–1987	Max and min temps	No strong singularity

From Godfrey et al., 2002, where references to cited articles are given.

The January Thaw, for example, is an anomalous warming that occurs in the northeastern United States in mid- to late January. As Table S5 indicates, this feature has been the topic of many studies. The conclusions are quite variable, with some indicating the existence of a singularity and others finding no consistent record.

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Cross-reference

Indian Summer

SNOW AND SNOW COVER

Importance of snow cover

Snow cover refers to the blanket of snow covering the ground, and includes the concepts of depth and areal extent (Sturm et al., 1995). Snow cover is a key component of the global climate system through its role in modifying energy and moisture fluxes between the surface and the atmosphere, and through its role as a water store in hydrological systems. Snow is a highly reflective material; the reflectivity or ***albedo*** of new snow is 0.8–0.9, which means that 80–90% of the incident solar energy is reflected away from the surface. This property, combined with the excellent insulating characteristics of a snow cover, dramatically reduces the energy exchange between the surface and the atmosphere. Empirical studies have shown that mean surface air temperatures are typically 5°C colder when a snow cover is present (Groisman and Davies, 2001). This positive feedback is responsible for the rapid expansion of northern hemisphere snow cover extent in October–November. The larger-scale climatic significance of the snow–albedo feedback is modulated by cloud cover and by the small amount of total solar radiation received in high latitudes during winter months. Groisman et al. (1994a) observed that snow cover exhibited the greatest influence on the Earth radiative balance in the spring (April to May) period when incoming solar radiation was greatest over snow-covered areas. The impact of snow on surface reflectivity is an example of a ***direct feedback*** to the climate system. Snow is also involved in a number of ***indirect feedbacks*** such as its influence on modulating sea ice growth, and linkages to cloud cover and surface temperature through snow's role in soil moisture recharge (Cess et al., 1991; Randall et al., 1994). There is an extensive body of literature linking snow cover feedbacks to monsoon circulations (e.g. Vernekar et al., 1995; Gutzler and Preston, 1997). However, recent research suggests snow's role may be limited (Robock et al., 2003). Groisman and Davies (2001) and Armstrong et al. (2003) provide more detailed reviews of the climatic significance of snow.

Snow cover is also a critical component in natural and human systems. Snow accumulation is an important resource for drinking water, irrigation, hydroelectrical generation and natural river ecosystems. The presence of an insulating snow cover is also critical for ecological systems (Jones et al., 2001). Beneath even 30 cm of snow organisms and soil are well protected from the extreme diurnal temperature fluctuations occurring at the snow surface, and exchanges of carbon, methane and other gases between the land surface and the atmosphere can continue during the winter period (Sommerfeld et al., 1993). Snow influences on soil temperature are also important for hydrology. When soil moisture freezes, the hydraulic conductivity is reduced, leading to either more runoff due to decreased infiltration or higher soil moisture content due to restricted drainage. Knowing whether the soil is frozen or not is important in predicting surface runoff and spring soil moisture reserves (Zhang and Armstrong, 2001).

Finally, snow supports a multi-billion-dollar recreation and tourism industry in midlatitudinal mountain regions of the world (e.g. Rocky Mountains, Appalachians, Alps). But snow can also be a nuisance and a hazard. Snow can play havoc with driving conditions through reduced visibility and traction, heavy loads of snow can collapse buildings, and rapid melt of snow can cause flooding. In 1987–1988 the snow removal budget for the city of Montréal was $47 million (Phillips, 1990). In mountainous areas of the world, snow avalanches are an ever-present hazard with the potential for loss of life, property damage and disruption of transportation. McClung and Schaerer (1993) provide an excellent review of avalanche science.

Snowfall

The initial characteristics of a ***snowpack*** start with the snowfall process. While the physics of snowfall formation is highly complex (see review by Schemenauer et al., 1981), the basic ingredients are a source of moisture, a mechanism for vertical motion in the atmosphere to cause precipitation to form (e.g. orographic lift, frontal lift, convection), and air temperatures at or below 0°C. The form of the initial ice crystals (e.g. columnar, platelike, dendritic) depends on the temperature at formation, and the crystals may undergo considerable change as they fall through layers with different temperature and humidity. The shape and size of the final snowflakes thus depend on a host of factors, which can be further complicated by wind action breaking crystals into smaller fragments. The density (mass per unit volume, expressed in units of $kg\,m^{-3}$) of natural snowfall varies with crystal type, size and liquid water content, but is typically in the range 50–120 $kg\,m^{-3}$ (Pomeroy and Gray, 1995). Snowfall densities vary from event to event, and there are important regional differences in snowfall densities based on temperature and precipitation climates. While some data suggest snowfall density is air temperature-dependent (e.g. Pomeroy and Gray, 1995), snowfall density is not strictly a function of air temperature, and densities can vary greatly for any given air temperature (Doesken and Judson, 1997). The density and crystal structure of newly fallen snow are important in many engineering and operational fields such as avalanche risk forecasting, design of roofs and structures to withstand snow loading, the accretion of snow on transmission lines (a particular problem in Japan where wet snow is frequent), aircraft deicing, and vehicle trafficability.

Spatial extent

Snow cover is encountered over most of the northern hemisphere mid- and high-latitudes during the winter season, and over many mountainous regions of the world for extended periods. Figure S21 shows the seasonal range in snow cover over the northern hemisphere from satellite data. The temporal variability is dominated by the seasonal cycle with average snow cover extent ranging from an average minimum extent of 3.6 million km^2 in August, to an average maximum extent of 46.8 million km^2 in late January (Table S6). Most of the global snow cover extent is located over the northern hemisphere: with the Antarctic and Greenland landmasses excluded, the southern hemisphere mean maximum terrestrial snow cover extent is less than 2% of the corresponding winter maximum snow cover extent over the northern hemisphere. In the southern hemisphere exclusive of Antarctica, most of the seasonal snow cover extent is located over South America.

While annual snow accumulations can exceed several meters in humid mountainous regions of the world, shallow snow covers in the order 10–30 cm depth are typical for large areas of the northern hemisphere that experience relatively cold, dry winters (e.g. midlatitudinal continental and polar regions). The contrast in snow depths between regions close to winter moisture sources (i.e. Pacific and Atlantic oceans) is evident in Figure S22, which shows mean February snow depths over North America derived from surface observations.

Snow cover characteristics

Over a winter season a snowpack grows and develops as a complex layered structure reflecting the weather and climate conditions of each precipitation event, and metamorphic and melt processes in the snowpack. The international standards (units, terminology, and symbols) for describing the vertical structure of a snowpack are provided in Colbeck et al. (1990). The three basic properties used to describe a snow cover are depth, density and snow water equivalent (SWE) related as follows:

$$\mathrm{SWE}\ (\mathrm{mm}) = 0.01\, h_s \times \rho_s$$

where h_s is the depth of snow (cm) and ρ_s is the density of snow ($kg\,m^{-3}$). The conversion from a mass of snow ($kg\,m^{-2}$) to a depth of water (mm) is based on the fact that 1 mm of water spread over an area of 1 m^2 weighs 1 kg. Over a snow season a typical snowpack is characterized by an extended accumulation period that may include discrete melt events, followed by a rapid melt or ablation period (see example in Figure S23). Snowpack

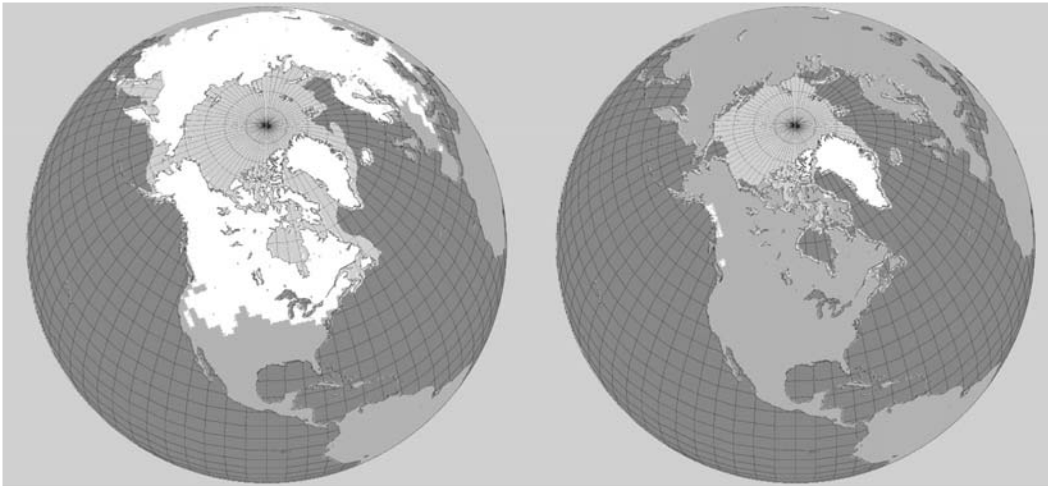

Figure S21 Mean seasonal variation in northern hemisphere snow (white) and sea ice (light grey) extent between February (left) and August (right) as derived from satellite data. Data are from Weekly Snow Cover and Sea Ice Extent, National Snow and Ice Data Center, 1996.

Table S6 Snow cover extent in million square kilometers for terrestrial portions of the globe (after Table 6.1 in Goodison et al., 1999)

Northern hemisphere	Excluding Greenland	Including Greenland
Maximum (late January)	44.8	46.8
Minimum (late August)	1.4	3.6
Southern hemisphere	Excluding Antarctica	Including Antarctica
Maximum (June–October)[b]	0.85[a]	14.5[c]
Minimum (January–March)[b]	0.07	13.7[c]
Global extent	Excluding Greenland + Antarctic	Including Greenland + Antarctic
Northern hemisphere winter	44.9	60.5
Southern hemisphere winter	2.3	18.1

[a] The main contribution is South America, which was estimated to have a mean winter snow cover extent of $0.45 \times 10^6\,km^2$ over the 1988–2001 period (Foster et al., 2003). New Zealand mean winter snow cover extent is estimated to be $0.06 \times 10^6\,km^2$ (Fitzharris and McAlevey, 1999).

[b] These month ranges are approximations since seasonal snow cover extent in southern hemisphere mountainous regions exhibits strong month-to-month and year-to-year variability.

[c] Seasonal snow cover variation in Antarctic snow cover was not taken into account.

density typically increases rapidly at the start of the season due to changes in the size, shape and bonding of snow crystals. This process is termed ***metamorphism***, and is caused by temperature and water vapor gradients, crystal settlement and wind packing (Pomeroy and Gray, 1995). During the melt season, snow density can also experience rapid increases from melt and refreezing. Melting snow densities typically range from 350 to 500 $kg\,m^{-3}$. In extremely cold environments with shallow snowpacks, snow density can actually decrease over time due to the formation of ***depth hoar*** in response to strong temperature and vapor gradients through the snowpack. A detailed review of the physics of snow metamorphism is provided by Langham (1981).

At a continental scale, differences in climate give rise to distinct snow cover–climate regions. Sturm et al. (1995) were able to classify global snow cover into six distinct classes with unique stratigraphic attributes (tundra, taiga, alpine, maritime, prairie and ephemeral) using wind, precipitation and air temperature data. Snow cover also exhibits extensive local scale variations due to the effects of wind redistribution, vegetation interception and sublimation, and the influence of topography on wind speed and the local energy balance. An extensive review of these processes is provided by Pomeroy and Gray (1995). Sublimation losses from blowing snow and tree canopies are important when determining the water budget of hydrological systems. Pomeroy and Gray (1995) estimated sublimation loss over prairie environments to be 15–41% of annual snowfall, and they estimated that approximately one-third of total snowfall falling on spruce and pine was lost through canopy sublimation.

Snowpack modelling

The ability to accurately model the accumulation and ablation of a snow cover is important for many applications such as flood forecasting, reservoir management, and climate system simulations. A wide range of approaches have been used to simulate the accumulation and ablation of a snowpack from simplified degree-day melt models to more complex physical models of the energy and mass balance such as Anderson (1976). In more

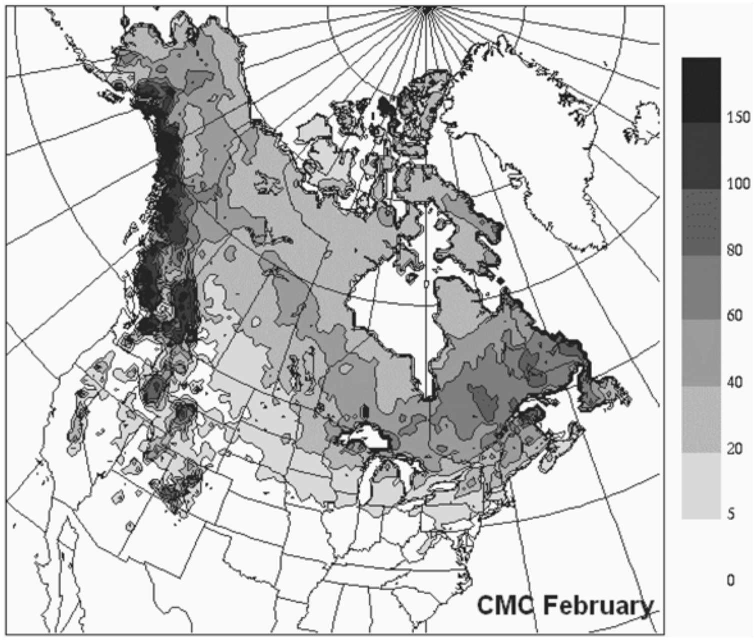

Figure S22 February mean snow depth (cm) over North America derived from an objective analysis of surface observations over the period 1979–1997 (Brown et al., 2003).

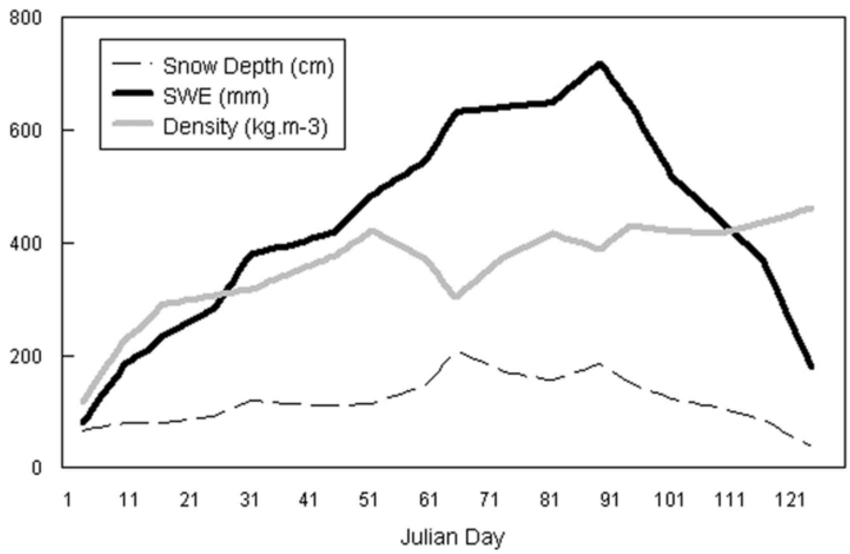

Figure S23 Temporal variation in snow depth (cm), SWE (mm) and mean snowpack density (kg. m^{-3}) from weekly snowpit observations at Col de Porte, France, 1995. Note the effect of the major snowfall event on Day 66 in decreasing the average snowpack density. Data courtesy E. Martin, Météo-France.

recent years, modelers have incorporated the physics of snow crystal size and shape evolution into physical models (e.g. Brun et al., 1992) along with improved understanding of heat transfer at the snow surface (Jordan et al., 1999), snowmelt (Marsh, 1999) and blowing snow (Pomeroy and Gray, 1995). The latest generation multilayer physical snowpack models have been shown to provide accurate simulations of point snowpack properties and melt at open locations for a range of sites and snow climates. Accurate simulation of snow cover over an area is a much more difficult problem, however, since this involves taking into account local-scale variations in topography and vegetation that are the main factors driving spatial variation in snow cover accumulation and melt. Important areas of ongoing snow modeling research include improved representation of blowing snow, vegetation and canopy processes, treatment of patchy snow, as well as development of approaches to include local-scale heterogeneity into global and regional climate models.

Measuring snow

The measurement of ***snowfall*** (the depth of freshly fallen snow that accumulates during the observing period, traditionally measured with a ruler), ***solid precipitation*** (the amount of liquid water in the snowfall intercepted by a precipitation gauge), and snow on the ground (depth, SWE) is a science in its own right. Reviews of measurement equipment and techniques are provided by WMO (1981), Goodison et al. (1981), Sevruk (1992), Pomeroy and Gray (1995), and Doesken and Judson (1997). Accurate information on the amount and solid fraction of winter precipitation is required for input to hydrological and climate models, but this is often difficult to provide in practice. For example, standard techniques have been developed for correcting precipitation gauges for wind-induced undercatch (Goodison et al., 1998) but not all observing sites have the required information to apply corrections. In addition, the available surface-based observational networks for snowfall and solid precipitation tend to be concentrated in populated lower-elevation areas, and in recent years many of these sites have been closed or converted to automated stations. The automation of precipitation measurements is difficult, and has important consequences for the homogeneity of climate data series.

Accurate information on the amount and spatial distribution of SWE is critical information for water resource management (e.g. agriculture, hydroelectric power generation, flood forecasting). Water authorities and utilities make use of surface-based snow surveys (and snow pillows in mountainous terrain) to monitor peak SWE values prior to snowmelt, as well as satellite imagery to map snow cover extent. Information on snow depth and SWE can also be derived from a variety of aircraft or satellite sensors with varying degrees of success (Hall and Martinec, 1985).

Temporal variability and climate change

Regional and continental snow cover exhibits large interannual variability in response to atmospheric circulation patterns that influence temperature and precipitation. Interannual variability in North American and Eurasian snow cover extent have been shown to be strongly correlated to the Pacific-North America (PNA) and North Atlantic Oscillation (NAO) circulation patterns respectively (Gutzler and Rosen, 1992). Snow cover exhibits large regional variability in response to ENSO events (Cayan, 1996; Clark et al., 2001), with a tendency for El-Niño events to be associated with greater snow cover over Eurasia and less snow cover over North America (Groisman et al., 1994b). A detailed review of snow cover–atmosphere relationships is provided by Groisman and Davies (2001).

Snow cover extent (SCE) exhibits significant negative correlations to air temperature in many regions of the northern hemisphere (see the “temperature response regions” in Groisman et al., 1994b), and for the hemisphere as a whole (Robinson and Dewey, 1990; Karl et al., 1993). This negative relationship (see Figure S24) reflects the positive snow–albedo feedback, and analysis of northern hemisphere snow cover has shown that the most significant decreases have occurred in the second half of the snow year when the snow–albedo feedback is strongest. Satellite records indicate that the northern hemisphere annual SCE has decreased by about 10% since 1966, largely due to decreases in spring and summer since the mid-1980s over both the Eurasian and American continents (Robinson, 1997, 1999). Analysis of reconstructed SCE information since 1915 (Brown, 2000) showed that most of the observed reduction occurred during the second half of the twentieth century. This period has been characterized by widespread trends toward less winter snow, earlier snowmelt and earlier snowmelt runoff, with important implications for water resources, e.g. reduced storage of water in the snowpack and earlier melt translate to a lower freshwater pulse for recharge of soil moisture and reservoirs, and increased potential for evaporation loss. Global Climate Model (GCM) simulations suggest widespread reductions in snow cover over the next 50–100 years in response to global warming. However, there is considerable uncertainty in model-projected regional patterns of snow cover change (Frei and Robinson, 1998), particularly in mountainous regions. A discussion of the impacts of reductions in snow cover is provided in Fitzharris (1996).

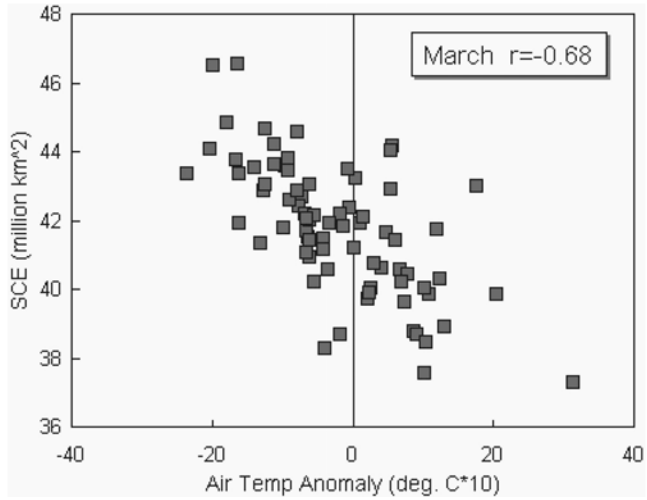

Figure S24 Scatterplot of reconstructed (1922–1971) and satellite-observed (1972–1997) northern hemisphere snow cover extent (SCE) versus northern hemisphere midlatitudinal (40–60°N) land surface air temperature anomalies for March. Air temperature anomalies were computed from the Jones (1994) gridded land temperature dataset, and the snow cover data are from Brown (2000). The inferred air temperature sensitivity for northern hemisphere March SCE is $-1.26 \times 10^6\,km^2\,°C^{-1}$.

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Cross-references

Albedo and Reflectivity
Antarctic Climates
Arctic Climates
Hydroclimatology
Taiga Climate
Tourism and Climate
Tundra Climate
Oscillations

SOLAR ACTIVITY

The sun is a variable star. A variety of transient phenomena are observed on or near its visible surface, including looping prominences and explosive flares. "Solar activity" is a general, inclusive term employed to characterize these and other phenomena, along with their variations in time. Some effects of solar activity are propagated throughout the solar system by the