OBSERVATION OF POLAR CLOUDS AND AEROSOLS FOR RADIATION BUDGET AND CLIMATE STUDY

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1. INTRODUCTION

Clouds and aerosols are believed to have a strong impact on climate through radiative effect; however, their 3-D distribution behaviors and interactions are still uncertain, especially in the polar regions.

In the present paper, recent status of the polar cloud climatology is reviewed first. Recently, decadal change in many aspects are found in the Arctic, and discussions are continued as for the indication of the long term trend. Cloud and radiation climatology is an crucial issue in these discussion, and perspectives of analyses and models are examined. Study on the polar cloud climatology is done for many years using satellite data. Extensive efforts have been devoted to the analysis of polar cloud detection from satellite passive measurements; however, cloud detection algorithm still have problems during long polar night when only the infrared channels could be used. Cloud detection using brightness temperature and brightness temperature difference in 10 mm split window channels of AVHRR was verified from the surface observation at inland of Antarctica.

Arctic aerosols are also discussed as another major radiative active components. A new airborne project on aerosol and radiation in the Arctic, ASTAR (Arctic Study on Tropospheric Aerosol and Radiation) 2000 is introduced. 3-D behavior and radiative effect of aerosols - Arctic haze - and then climate impact are to be derived.

Following these basic study on aerosols and clouds in the both polar regions, it is expected to have an active measurements by lidar and radar, together with passive imager and FTIR, from satellite in order to derive 3-D distributions under the future plan of ATMOS-B1 mission.

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2.POLAR CLOUD CLIMATOLOGY

The polar regions are receiving increased attention in climate change scenarios obtained from climate models. These attention arises from the "polar amplification" of the surface warming in simulations with increased concentrations of greenhouse gases. The zonally averaged warming in the Arctic is two or three times larger than the global warming for a doubling of CO2 (Randall et al., 1998; Houghton et al., 1990). The enhanced warming in high latitudes is partly attributable to the retreat of sea ice, and this retreat is accelerated by the positive feedback between temperature and surface albedo (Walsh and Chapman, 1998).

In the recent, large variation in many climate parameters are found in the Arctic as "Arctic Change". Warming of the air temperature near the surface is clearly shown in the real world, especially over the continent, north west of Canada and Alaska, and Siberia (Chapman and Walsh, 1993). The maximum trend was 0.75 °C/decade, and the relation to the regional trend of sea ice extent was discussed. The change in sea ice extent in the whole Arctic was reported by Cavalieri et al. (1997) as to be slight negative, -2.9 %/ decade, based on long discussion as for the continuity of the sensitivity of satellite microwave radiometers.

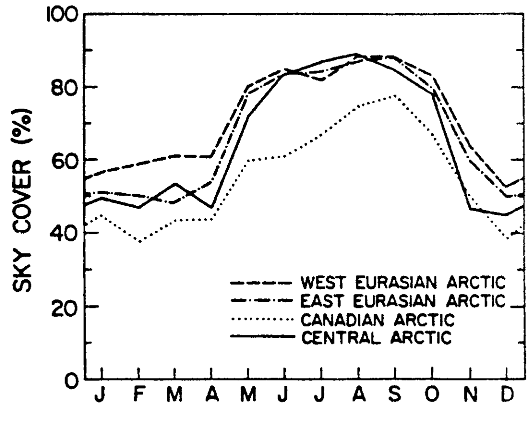

Fig. 1 Monthly mean Arctic cloud cover (from Barry et al., 1987, after Huschke, 1969).

The validity of the large polar amplification is confronted by several issues, one of which is associated changes in polar cloudiness (Stone, 1997). From the historical observations, cloud cover in the Arctic are known to have a distinct seasonal variation as seen in Fig. 1. Cloud amount about 50 % continues through winter, then increases abruptly in May, and then higher cloud amount continues during summer. However, polar clouds are generally not well simulated by global climate models (Walsh and Chapman, 1998). Tao et al. (1996) show that even the cloud fractions vary tremendously among the global climate models used in the Atmospheric Model Intercomparison Project as seen in Fig. 2.

The results obtained from the observational data was compared by Walsh and Chapman (1998) with corresponding evaluations from two recently released sets of atmospheric reanalysis products, from two major centers; NCEP (National Center for Environmental Prediction) and ECMWF (European Centre for Medium-Range Weather Forecasts). The observational data are from the Russian ice station, "North Pole" series NP2 to NP31 during 1950 to 1991 (Marshunova and Mishin, 1994). While global climate models are the final targets of the study, the reanalysis products represent an attractive intermediate vehicle for assessment, because the models used to assimilate data for the reanalyses are almost the same as the models used to simulate global climate, and the observational constraints on the reanalyses can be expected to result in more realistic depiction of the primary atmospheric variables. First of all, the distribution of Arctic cloudiness are striking different in the NCEP and ECMWF reanalysis. The NCEP cloud fractions are nearly identical in summer and winter. By contrast, the ECMWF distribution shows a strong bimodality, and its seasonality shows much closer correspondence with the observations. Then Fig. 3 summarizes the maximum cloud radiative forcing from observation and analyses. The observational values from the AARI data are positive in all months from September through April and negative from May through July. Neither reanalysis captures the essential characteristics of this seasonal cycle: the NCEP values remain positive in all months, although they show similar pattern of the seasonal variation as AARI data; the ECMWF values are close to zero in all months and show no seasonality, primarily because clouds do not impact the surface solar radiation in ECMWF values. Present reanalyses are less successful in capturing the annual cycle of Arctic clouds.

By the way, observation it self of cloud amount still has problems and we do not have any reliable cloud climatology in the Arctic and Antarctic. Especially satellite cloud climatologies in the polar region accompany difficulties, as seen in Fig. 4 (Rossow et al., 1993). Results of International Satellite Cloud Climatology Project (ISCCP) shows a opposite seasonal pattern with the surface observation both in the Arctic and Antarctic. There are several attempts to improve global cloud climatology with special attention to the polar region.

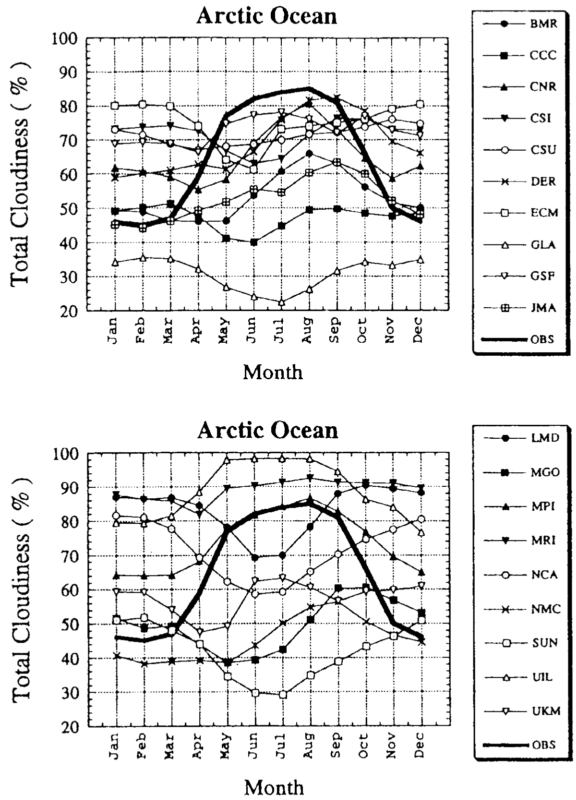

Fig. 2 Annual cycle of monthly mean total cloudiness (%) over the Arctic Ocean as simulated by the 19 AMIP models. Heavy lines show the annual cycle for 75 - 85° N derived from observations made at drifting ice stations (Tao et al., 1996).

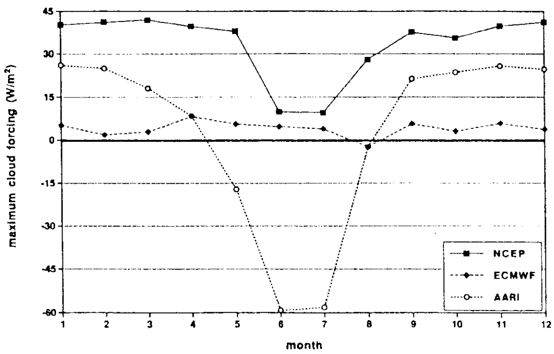

Fig. 3 Annual cycles of maximum cloud radiative forcing from the NCEP reanalysis (solid squares), the ECMWF reanalysis (Solid diamonds) and AARI ice station data (open circles) (Walsh and Chapman, 1998).

Recently, the reprocessed ISCCP DX data (Rossow and Shiffer, 1999) and data from revised TOVS Improved Initialization Inversion (3I) algorithms are compared (Stubenrauch et al., 1999). The former shows rather reasonable amount, while the latter shows extremely high cloud amount over the Antarctic continent.

3. VALIDATION OF SATELLITE CLOUD ALGORITHM IN THE ANTARCTIC INLAND AREA

As shown in the previous section, polar cloud climatology from satellite data is still uncertain. In the polar region over snow covered surface, simple cloud detection algorithm of visible or infrared data will confronted with difficulties due to high albedo and low temperature of snow surface, which makes the contrast between cloud and snow surface small. Extensive efforts have been devoted to the analysis of polar cloud detection from satellite passive measurements (e.g.., Raschke, 1987; Lubin et al., 1997); however, cloud detection algorithm still have problems during long polar night when only the infrared channels could be used. Cloud detection using brightness temperature and brightness temperature difference in 10 μm split window channels of AVHRR (Yamanouchi et al., 1987; Yamanouchi and Kawaguchi, 1992) was validated from the surface observation at Dome Fuji Station (77°19' S, 39°42' E, 3800 m a. s. l.; Fig. 5) at inland of Antarctica. We had a very precious chance of wintering observation for atmospheric science at Dome Fuji Station in 1997, following two years of wintering for deep ice core drilling (Fujii et al., 1999). Though the data of channel 3 of AVHRR at 3.7 μm should have an information of clouds even without solar reflection, they contains large variabilities up to 40 °C at low temperatures such as -80 and -90 °C due to low temperature resolution, and are difficult to use.

As shown in Fig. 3 of Yamanouchi and Kawaguchi (1992) as the successful example, we tried to make a scatter plot of those data for the pixel covering the Dome Fuji Station. Pixels of 128 x 128 with 2.2 km resolution are sampled from about 300 km square and 4 x 4 pixels are averaged. Then data of 32 x 32 averaged pixels are shown in Fig. 6. Typical results from 9 days (9 passes) in June 1997 are shown in this figure, data of 2, 11 and 25 when the surface observer reported to be clear and of 17 and 18 when the surface reports are overcast with blizzard, and also of 19, 20, 23 and 24 when the sky is covered with cloud amount larger than 8. In case of clear sky, data points distributed around T4 - T5 = 0, with the range of T4 between -85 to -60 °C. On 17 and 18, T4 ranges between -75 and -55 °C, and T4 - T5 distributed widely between 2 and 6 °C, easily detectable as cloudy. On other cloudy days, 19 to 24, T4 - T5 still show some distinction from 0 °C and detectable as cloudy. These data all show a positive possibilities of cloud detection by T4 - T5.

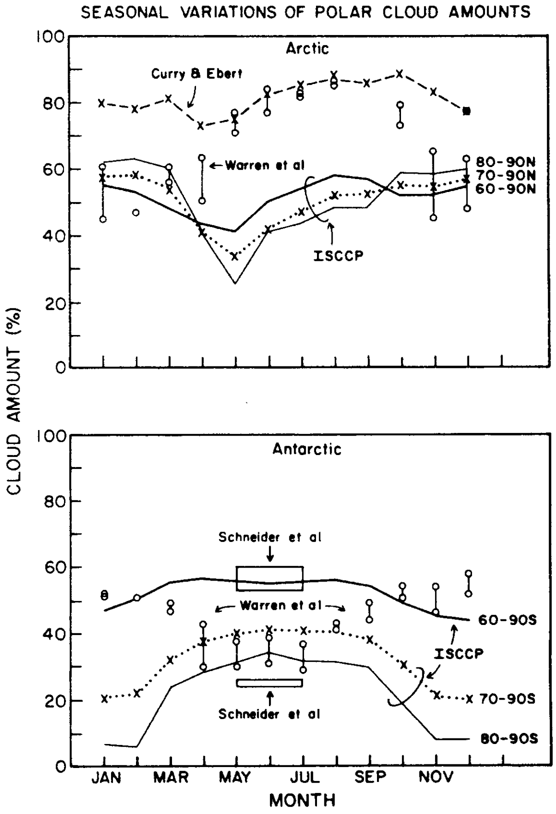

Fig. 4 Comparison of the seasonal variations of average cloud amount in the Arctic and Antarctic determined by ISCCP and other cloud climatologies (Rossow et al., 1993).

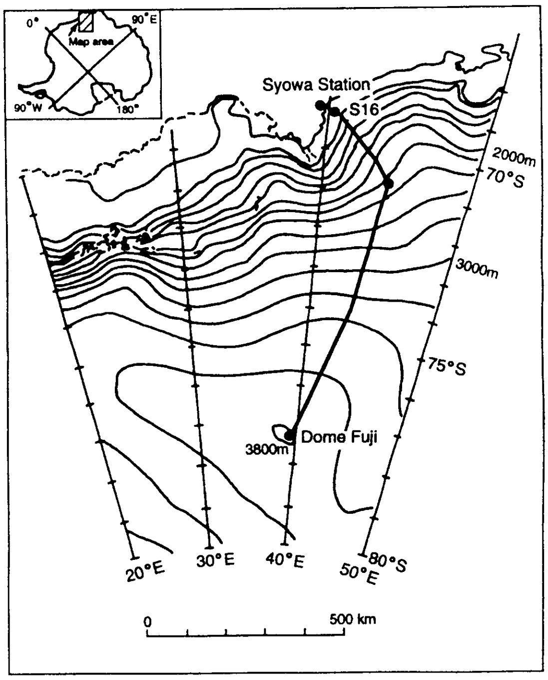

Fig. 5 Location of Dome Fuji Station inland of Antarctica.

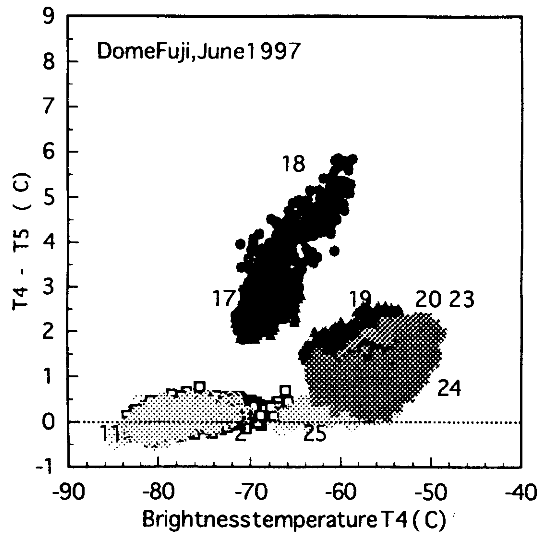

Fig. 6 Comparison of scatter diagram of T4 against T4 - T5, Dome Fuji Station, June 1997.

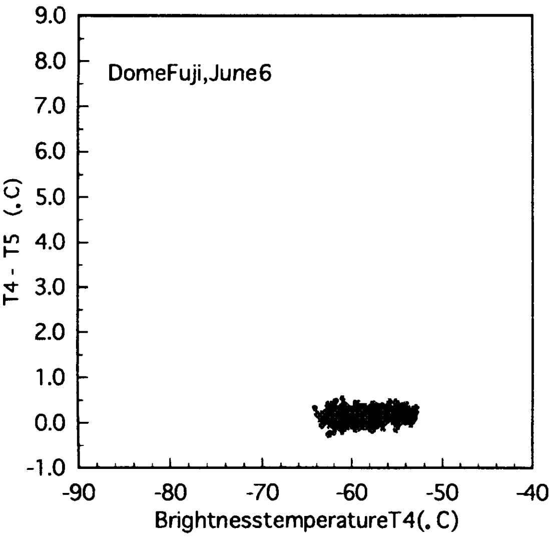

Fig. 7 Scatter diagram of T4 against T4 - T5, with low level cloud, Dome Fuji Station, June 6, 1997.

On the other hand, Fig. 7 shows the similar scatter plots for July 6, when the surface reports were cloud amount 8 with Cs. Examining the results of longwave radiation measurements as shown in Fig. 8, increased downward radiation assured the cloudy condition on this day. From Fig. 7, data points lie along T4 - T5 = 0, and are difficult to know the difference with clear cases as seen in Fig. 6. However, the range of T4 is very high about -65 to -50 °C, which might be some indication of cloudy condition.

Fig. 9 shows the average T4, average (T4 - T5) and cloud amount reported by the surface observers at the time of satellite pass. As mentioned above, most of cloudy skies make the T4 - T5 large and become discriminatable with clear skies. However, in some case, such as 6 and 15 June, T4 - T5 is nearly 0 and difficult to distinguish from clear case. However, on these cases, T4 itself shows an abrupt increase from a day before, and from the variation of T4, it is possible to say as cloudy. On these case, most of clouds are low level and stratiform, and lie in the high temperature layer over the surface inversion. Once clouds cover the area, downward radiation increases and also increase the surface temperature and upward radiation (Fig. 8). Large T4 does not always mean to be cloudy. Since, once cloudy sky continues and air temperature increases, the surface temperature does not decrease so quickly and continues to be higher even in the clear sky. Only the abrupt increase of T4 is acceptable to be brought by cloudiness. Consequently, most case of cloudy skies, at least all cases with cloud amount larger than 5 in June 1997, were detected by AVHRR T4 and T4 - T5.

Return to the variation of longwave radiation flux shown in Fig. 8. The base line for the downward longwave flux is about 70 to 80 W/m2, which is for clear day. Increase of about 30 to 50 W/m2 is frequently seen, which is due to the covering by clouds. However, sometimes as on 17 to 18, there is a tremendous increase of downward flux up to 200 W/m2. Upward flux also follows the similar increase. This great increase is not only caused by the cloud radiative effect, but also caused by the effect of "blocking condition". Explanation of this phenomena in detail is already done by Hirasawa et al. (2000). Once the blocking high is settled, warm and moist air is liable to advected into high plateau of inland Antarctica, and surface air temperature increases greatly. In the present case, it was more than 40 °C! From aerological observations made at the station (Hirasawa et al., 1999), temperature and humidity profile changes greatly and absolutely different atmosphere such as mid-latitude air is intruding into the Antarctic. This great change in the airmass makes the radiation change greatly. The role of this phenomena to the energy transfer of the polar regions is a matter of discussion.

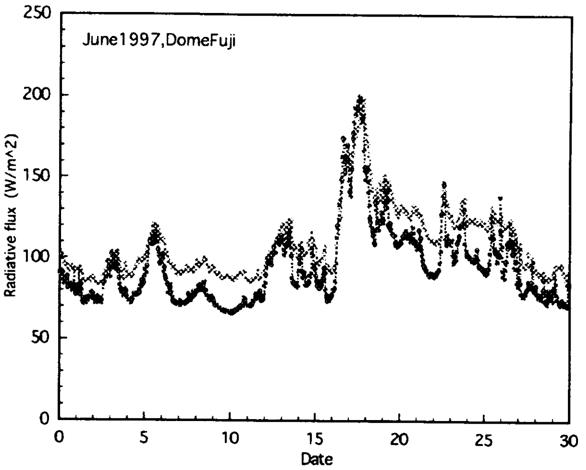

Fig. 8 Longwave radiative fluxes at Dome Fuji Station, June 1997.

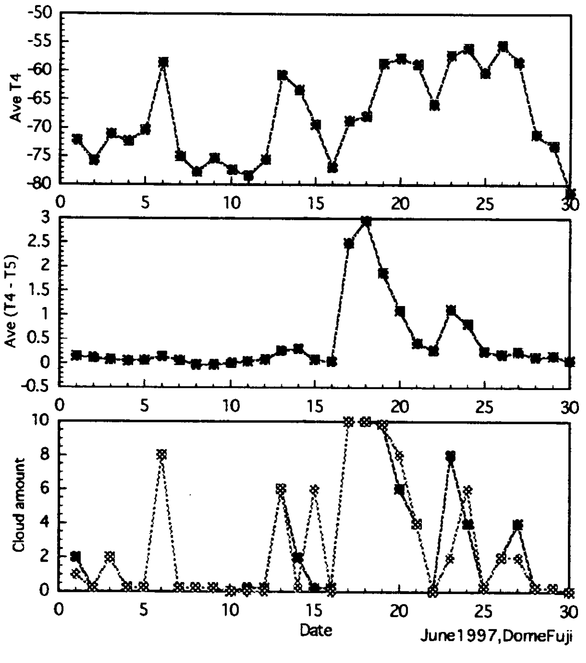

Fig. 9 Average T4, average (T4 - T5) from AVHRR around Dome Fuji Station and cloud amount reported by the surface observers at the time of satellite pass.

4. ARCTIC AEROSOLS

Aerosols are another constituents which affect radiation budget and climate. In the Antarctic, base line of the optical depth of aerosols is about 0.02 (at 500 nm; Shaw, 1988), and stable except under the effect of volcano eruption (Herber et al., 1996). In the Arctic, baseline is still higher as 0.05 and optical depth shows a large seasonal variation. Fig. 10 shows the seasonal variation of optical depth measured by sun, moon and star photometers at Ny-Alesund, Svalbard (Herber, private communications). From the late winter to spring, optical depth increase greatly, up to 0.2. This large spring maximum of aerosol loading is known as "Arctic Haze", originated from surrounding anthropogenic sources, and trapped long time in the strong inversion layer with lacking any effective removal process (Shaw, 1995). Though this is only a seasonal phenomena, its radiative effect is not negligible, because the solar radiation becomes effective just in this season after long polar night and large influence will be expected related to the melting and decay of sea ice. Many observational studies have been devoted to the Arctic haze till the recent; however, not so much were done for the radiative effect yet.

A new Japan-German cooperative project on aerosols in the Arctic, ASTAR (Arctic Study on Tropospheric Aerosol and Radiation) 2000 is to be carried out in coming March and April. Airborne observations of vertical distribution of physical, chemical and optical properties of aerosols, Arctic haze, will be made around Svalbard using German aircraft Polar 4 (Dornier 228) of Alfred-Wegener Institute for Polar and Marine Research. Also, remote sensing of aerosols, in situ measurements and sampling of aerosols will be made coordinately at the surface of Ny-Alesund Scientific Station, Svalbard (Fig. 11). At this area, scientists from several countries including German and Japanese scientist (Yamanouchi et al. 1996), have already conducted ground based atmospheric science observation for

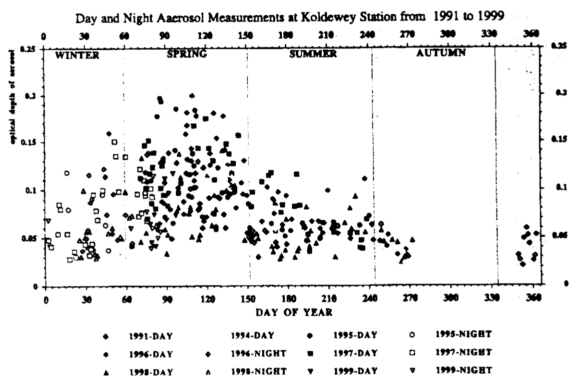

Fig. 10 Day and night aerosol optical depth measured at Ny-Alesund, Svalbard from 1991 to 1999 (Herber, private communications).

these 10 years. It is a good site to have an international cooperative campaign. Data are compared with the satellite SAGE-II measurements. Also, data will be used as an input for the regional climate model to estimate the radiative effect. 3-D behavior and radiative effect of aerosols - Arctic haze - and then climate impact will be examined.

The sensors to be launched on the aircraft are sun photometer, broad band radiometers (up and down facing), particle soot absorption photometer (PSAP), integrating nephelometer (IN), optical particle counter (OPC), aerosol impactor and aerosol sampler, together with basic meteorological instruments (Fig. 12). The flight will be conducted from Longyearbyen, Svalbard, for about 15 flights with 75 hours. Major instruments for the ground based observations are Raman lidar, micropulse lidar, sun and star photometers, sky radiometer, broadband radiometers, radiosonde, aerosol sonde (OPC), optical particle counters, aerosol samplers, microwave radiometers, radar and POSS. Together with these observations, continuous observations on greenhouse gases, ozone and other trace species are also conducted at the station. Members from Alfred Wegener Institute for Marine and Polar Research, Norwegian Polar Institute, Norwegian Atmospheric Research Institute (NILU), Meteorological Institute, Stockholm University (MISU), NASA Langley Research Center, Hokkaido University, Nagoya University and National Institute of Polar Research will join the project.

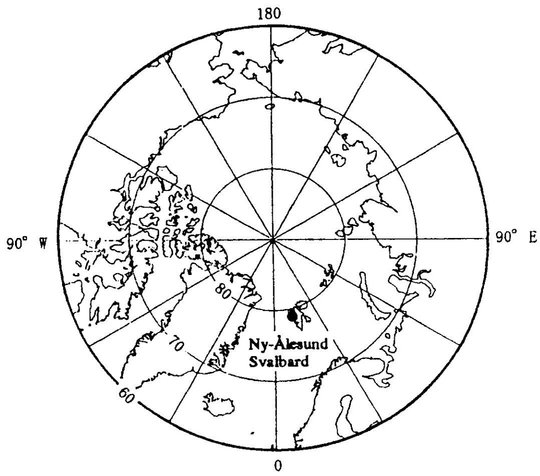

Fig. 11 Location map of Ny-Alesund, Svalbard in the Arctic.

5. CONCLUSION

Polar clouds are playing a key role in the climate system of the polar regions, and then in the global climate. However, even their climatology is not yet ascertain and further study is still needed in observations, analyses and modeling. Satellite cloud algorithm in the polar regions still has uncertainties, especially for the inland of Antarctica. Cloud detection using the brightness temperatures of split window channel in 10 μm region shows a positive perspective; however, not yet complete. A new observation project is to be conducted on the radiative properties of Arctic aerosols; however, the results will be limited on the time and space. It is concluded that global observation of the three dimensional distribution of clouds, aerosols and radiation by cloud profiling radar and backscatter lidar on a satellite is indispensable (3D-CLARE; ATMOS-B1).

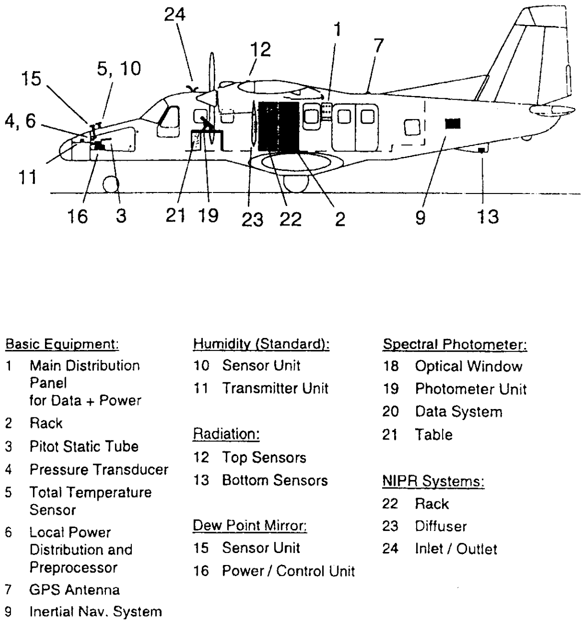

Fig. 12 Instruments on Polar 4 for ASTAR 2000.

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