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REPORT OF THE PICES-STA WORKSHOP ON MONITORING SUBARCTIC NORTH PACIFIC VARIABILITY

Jointly sponsored by PICES and the Science and Technology Agency of Japan

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Co-Chairmen:
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FOREWORD

The PICES Science Board and the Science and Technology Agency of Japan held a Workshop on Monitoring Subarctic North Pacific Variability, October 22-23, 1994, in Nemuro, Hokkaido, Japan, in conjunction with the PICES Third Annual Meeting. The Workshop addressed the responses of the subarctic Pacific to forcing by climate variations and human activities. The time scales considered were seasons to centuries. The first workshop objective was to suggest current and future long-term monitoring programs to describe significant forcing and responses. Second was to suggest scientific, technological and other factors that affect the above monitoring programs, and third, was to provide a summary report that advised on a strategy for developing a monitoring program, by 31 January, 1995. A preliminary report and recommendations were to be provided to the Science Board at the conclusion of the Workshop.

The Workshop was not intended to discuss process studies or to review the science of the subarctic Pacific, but rather to focus on the long-term monitoring programs required for assessment of the physical and ecological responses to long-term forcing, both natural and man-made. It was expected that much of the monitoring might occur under the auspices of the Global Ocean Observing System and the Global Climate Observing System.

The Workshop was structured around two themes and several sub-themes:

Subarctic Pacific Climate Variability
- External forcing; Global Climate Change
- Subarctic North Pacific
- Ice-Covered Marginal Sea
- Space and Time Scales; Sampling Problem

Ecosystem Response
- Subarctic Large-Scale Variability
- Lower Tropic Levels
- Higher Tropic Levels

Workshop Convenor:
- Y. Hayashi, Science and Technology Agency, Japan.

Steering Committee Co-Chairmen:
- Y. Sugimori, Tokai University / Chiba University, Japan.
- M.G. Briscoe, NOAA, U.S.A.

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- S. Ishii, JAMSTEC, Japan
- W.S. Wooster, PICES

Observers:
- H.D. Ahn, KORDI, Korea
- A. Bychkov, POI, Russia
- V.V. Navrotsky, POI, Russia

AGENDA

October, 22 (Saturday):

1. Introduction

2. Opening remarks by W.S. Wooster

3. Keynote address by Y. Hayashi, Y. Sugimori and M.G. Briscoe

4. Subarctic Pacific climate variability
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   ii) **L.D. Talley and X.J. Yuan**. Summary of variability of physical conditions in the subarctic North Pacific
   iii) **V.V. Plotnikov and G.I. Yurasov**. Seasonal and interannual variability of ice cover in the North Pacific marginal seas
   iv) **S.C. Riser**. Space and time scales of variability in the subarctic North Pacific: Implications to monitoring the system

5. Ecosystem Response
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   ii) **H. Ishii, and T. Takizawa**. Long-term variations of atmosphere and ocean in the PICES region
   iii) **S. Saitoh, H. Onishi, and M. Onishi**. An applicability of global multichannel sea surface temperatures to study on transition zone in the central North Pacific Ocean
   iv) **A. Harashima, R. Tsuda, Y. Tanaka, T. Kimoto, H. Tatsuta, and T. Hagiwara**. Ferry-based biogeochemical monitoring from the marginal seas to the subtropical gyre
   v) **Y. Senga, T. Araga and Y. Sugimori**. Modularized real-time surface-water monitoring system
   vi) **N. Kimura, Y. Okada and H. Fukushima**. Estimation of vertical profile of chlorophyll concentration around the Antarctic Peninsula derived from the satellite image (Nimbus-7/CZCS)
   vii) **M. Ikeda, T. Yao and Q. Yao**. Seasonal evolution of sea ice cover and shelf water off Labrador simulated in a coupled ice-ocean model
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INTRODUCTION: OCEANIC INTERDECADAL CLIMATE IN THE NORTH PACIFIC OCEAN AND LIVING RESOURCES VARIABILITY

1. INTRODUCTION

Although there are many important examples of climate fluctuations with quasi-periods of 10-30 years, such interdecadal climate variability (ICV) has been given relatively little attention by scientists, particularly in comparison with the 3-5 year time scale associated with the El Nino Southern Oscillation (ENSO). Instances of ICV are found in many physical and biological systems and can be discerned in high-resolution proxy records that are unlikely to have been affected by humankind.

Dynamically, the oceans are likely to play a key role in setting the time scale for ICVs. Besides having important socio-economic effects, an understanding of natural decadal-scale climate variations is crucial to differentiating natural climate changes from those due to anthropogenic forcing, e.g., Greenhouse Warming. This class of interdecadal variations is particularly challenging because as well as the oceans, it involves interactions between the atmosphere, ocean and cryosphere, and is probably also related to biogeochemical process.

Observations of global mean air temperature reveal year-to-year variations superposed on a long-term trend throughout the 20th century. However, there is also important decadal-scale variation (Nitta and Yoshinari, 1992). The warmest of the past decade is part of a long term trend of global warming associated with the observed greenhouse gas increases in the atmosphere (IPCC 1990).

A notable climate impact is the influence on the change in the pattern of migration of salmon to the mouth of Fraser River from the south end of Vancouver Island to the north end. Other connections range from an increase in the chlorophyll concentration of the surface water over the North Pacific gyre.

Another recent example of ICV has come from the North Atlantic. The remarkable phenomenon is well known as "Great Salinity Anomaly (GSA)". This event involves a freshening of the surface waters of the northern North Atlantic during the 1960s and 1970s. On the other hand, deep water masses of the world ocean form at both polar extremes of the Atlantic. In contrast, the Pacific has a circulation more constrained to horizontal layers. A freshening such as the GSA may interrupt the northward transport of heat in the Atlantic. In the North Atlantic, deep water formation would reduce poleward heat transport by the thermohaline circulation. Aside from redistributing the earth's heat and salt, the thermohaline circulation mixes water mass properties. Convective water mass formation also plays an important role in providing a sink for atmosphere CO_2 and the global CO_2 budget in the North Atlantic. Normally, information on changes below the ocean surface are scarce. In the North Pacific, there is evidence that the cooling in SST's over the recent decades extends to well below 500m, and involves a considerable amount of heat capacity.

In the remainder of this introduction, we focus on examples of interdecadal climate variability (the mid-1970s North Pacific climate shift) to give a perspective of what is known and what is unknown about this class of climate variability.

2. MID 1970S NORTH PACIFIC CLIMATE SHIFT

A reliable analyses of regional patterns of sea surface temperature (SST) anomalies for decades over the globe can only be carried out for the period after 1946. The most striking and largest amplitude anomalies of decade-mean temperature in the Northern Hemisphere are found for 1977-1986. In that decade, substantially, cooler (>0.5°C) than normal conditions (relative to the 1951-80 mean) occurred in central North Pacific SST. There was a large warming (>1.5°C) over Alaska and SST increased along the west coast of North America, while cooling occurred over eastern North America. This distinctive pattern of surface temperature anomalies for 1977-

86 is linked to changes in many other climate variables. At the North Pacific surface, the wind stress and wind stress curl changed, and by implication the Ekman pump, Sverdrup transport and ocean current. Sea surface temperatures, air temperature at the sea interface, sea-ice, and stream flow in coastal regions all changed. Other evidence suggests pronounced changes in phytoplankton, ocean biology and in fish stocks in the North Pacific. The changes appear to be closely linked with low frequency variability in the tropical Pacific and Indian Ocean and the El Nino Southern Oscillation (ENSO) phenomenon. At the ocean surface, the wind changes drive variation in surface fluxes, ocean mixing and ocean current. In the sea of Okhotsk, sea-ice cover often shows a reverse tendency with greater than normal values after about 1976 (Sekine 1989). Large changes are also found throughout the troposphere, that is, the pattern associated with strong changes downstream over North America. This "Pacific-North American" teleconnection pattern is closely linked to the changes in surface temperature, moisture and involves changes in cloud cover.

Meanwhile, over the Pacific Northwest, the shift to a stronger high pressure ridge over western Canada provided changes in both temperature and precipitation. Consequently, air temperature and sea-ice cover along the coast of Alaska changed (Salmon 1992). Also stream flow in many rivers along coastal Alaska increased (Cayan 1991).

Changes in the atmospheric circulation alter the sensible and latent heat fluxes from the ocean surface into the atmosphere. The flux changes observed during the 1970s and 1980s are consistent with changes in the observed SST pattern (Cayan 1992, a,b,c). A basin wide change in winds is apparent due to the difference between wind stress of 1977-1982 and 1971-1976 (Cayan et al.) with stronger cyclonic circulation in south of the Aleutian Islands after 1976. In the North Pacific, these changes imply changes in the curl of wind stress which has implications for the Ekman pump and Sverdrup transport (Salmon 1992).

The decadal changes during the mid 1970s also had profound effects on the central and eastern Pacific epipelagic ecosystem (Ebbesmeyer et al 1991). Measurements north of Hawaii suggest an increase in total chlorophyll in the water column and thus, in phytoplankton (Venrick et al 1987). The changes in ocean currents and temperatures around 1976 have also evidently altered the migration pattern of fish, in particular tuna and salmon in the North Pacific (Bernet and Mysak 1986).

The space-time scales of the dominant ENSO climate signal in the Pacific has been associated with a greatly reduced phytoplankton biomass in both the western (Dan Donneau 1986) and eastern Pacific (Barber et al 1983).

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Proceedings:
SUBARCTIC PACIFIC CLIMATE VARIABILITY

Long-term Variations of the North Pacific

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ABSTRACT

Long-term variations of sea surface temperature (SST), atmospheric forcing of the oceanic circulation and the upper thermal condition in the North Pacific are briefly described. It is insisted that the subsurface thermal data taken by TRANSPAC program are very useful to detect the interannual to decadal variations of the upper ocean thermal conditions and it is proposed that new monitoring program should be established to replace the TRANSPAC program.

1. INTRODUCTION

It is urgently necessary that we clarify the role of the ocean in global climate change due to the increasing greenhouse gasses. To do this, it goes without saying that a careful search for various kinds of archived oceanic data is needed. Since interannual to decadal scale variations of climate are considered to be major manifestations of large-scale air-sea interaction, special attention must be paid to both atmospheric and oceanic conditions and their mutual relationship. El Nino/Southern Oscillation (ENSO) events are good examples of such phenomena. Recently, many authors have pointed out significant signals with decadal to interdecadal time scales in the North Pacific and in the Northern Hemispheric atmospheric circulation. These phenomena are also considered to be the manifestations of large-scale air-sea interaction and to be the next investigation target of ENSO events.

In this paper, recent studies of long-term variations of sea surface temperature (SST) fields, atmospheric forcing of the oceanic circulation and the thermal conditions of the upper ocean will briefly be reviewed. In addition, it will be shown that subsurface thermal data taken by the TRANSPAC program have been very useful, and indeed critical, in the detection of long-term variations of the upper thermal conditions in the northern North Pacific.

The remainder of this paper is organized as follows: in Section 2, long-term variations found in SST fields of the North Pacific are reviewed and in Section 3, and changes of the atmospheric forcing of the oceanic circulation are also described. In Section 4, preliminary result of change of the upper ocean thermal condition is briefly described and in Section 5, the necessity of a new monitoring program is stressed.

2. LONG-TERM VARIATIONS OF NORTH PACIFIC SST FIELDS

Empirical orthogonal function (EOF) analyses for SST fields have been completed by several authors since late 1970s. Recently, Tanimoto et al. (1995a) have shown the existence of two distinctive patterns: one is that with the El Nino/Southern Oscillation (ENSO) time scale of 3-4 years (ENSO mode: Fig. 1). The other is that with decadal (DC) time scale (decadal mode: Fig. 2). The ENSO and decadal mode patterns were extracted by examining the band of periods 24-60 months, and the low frequency band, longer than 60 months, all computed from winter SST anomalies.

The leading mode of the DC time scale is a meridional dipole pattern: a north-south oscillation of low-latitude and mid-latitude oceans. Time coefficients of this mode (Fig. 2(b)) showed abrupt changes, like a step function, around 1976, i.e., mid 1970s, and continued to late 1980s. This abrupt change may be called a "Climatic Jump" or "Regime Shift" and will be discussed in the next section. In addition to this abrupt change, we can observe relatively short time-scale fluctuations in the 1950s.

This DC mode pattern is also clearly extracted by the composite analysis for the two categorized winters (DP and DN winters, see figure caption of Fig. 2 for the definition), as shown in Fig. 3. This means that the pattern extracted by EOF analysis is well organized and is a very robust feature.

It was also found that wind stress fields over the North Pacific varies coherently with SST fields described above, as shown in Fig. 4, in which composite maps of wintertime wind stress vector anomalies are shown (Tanimoto et al., 1995b). In DN (DP) winters, the elongated elliptical cyclonic (anticyclonic) pattern centered on 40N and the international date line appeared in the anomaly fields. These anomaly fields of wind stress vectors correspond to the southward (northward) shift of the westerly axis and strengthening (weakening) of westerlies in DN (DP) winters.

3. ABRUPT CHANGE OF ATMOSPHERIC FORCING OCCURRED IN MID 1970S

The abrupt change of atmospheric circulation over the North Pacific occurred around the mid 1970s and was noticed in late 1980s by meteorologists: e.g., Kashiwabara (1987), Nitta and Yamada (1989) and Trenberth (1990). Nitta and Yamada (1989) showed that since the mid 1970s, SST anomalies in the whole equatorial region of the Pacific Ocean have taken positive values, like a condition during ENSO events (see Fig. 3(b)). These positive SST anomalies in the equatorial region forced excitation of standing Rossby wave trains, the so-called Pacific/North American (PNA) teleconnection pattern over the North Pacific, which is responsible for the strengthening and eastward shift of the Aleutian Low, as well as strengthening of wintertime westerlies over the Pacific sector.

It is easily imagined that the strengthening of wintertime midlatitude westerlies since the mid 1970s would change the forcing of oceanic conditions in various aspects. The lowering of SSTs in the central Pacific as extracted by EOF analysis mentioned above. can be interpreted as one of the results. The strengthening of wintertime midlatitude westerlies can also cause an increase of Sverdrup transports in both the subtropical and subpolar gyres as shown by Hanawa (1995). Actual evidence of the spin-up of the subtropical gyre will be described in the next section. In addition, it is very interesting to note that since the mid 1970s to late 1980s, the Oyashio First Intrusion gradually penetrated southward off the Sanriku Coast (Hanawa, 1995) and the large meander of the Kuroshio off Honshu Island occurred since 1975.

4. CHANGE OF SUBSURFACE THERMAL CONDITION

Analysis for the upper thermal condition also shows the existence of DC time scale variation coherent with SST fields mentioned above. Watanabe and Mizuno (1994) analyzed the subsurface thermal data including XBT (Expendable Bathythermograph) data taken in the North Pacific. Fig. 5 shows a time-longitude matrix of 0-400m vertically averaged temperature anomalies along the 35-45°N latitude band (a), and that of SST (b). It is seen that heat content anomaly changed its sign from positive before 1976 to negative after 1976, corresponding to that of SST anomalies.

Spin-up/spin-down of the North Pacific subtropical gyre in recent decades could also be detected by using the subsurface temperature data (Yasuda and Hanawa, 1995). Fig. 6 shows the Sverdrup transport at 30°N and 130°E (a) and the temporal change of heat content difference between the north (39°N) and south (29°N) regions of the Kuroshio Extension. Although there is some time lag between both time series, they clearly show the increasing trend from 1960s to 1980s. This means that the North Pacific subtropical gyre has gradually been spun-up during the analyzed period. This change can be explained by that of wind stress field over the North Pacific (see Fig. 4). It was also shown that, corresponding to this change, thicknesses of the wintertime mixed layer in the midlatitude ocean and those of 15-19°C layer (Subtropical Mode Water) in the northwestern subtropical gyre vary coherently.

Analysis for the long-term variations of the upper ocean thermal conditions has just begun. Careful inspection of those data will give us much interesting and useful information on the oceanic role in climate change.

5. CONCLUDING REMARKS - ESTABLISHMENT OF NEW MONITORING PROGRAMS IN THE PICES AREA

As mentioned in the previous section, the subsurface thermal data are essential for clarification of the change of oceanic conditions, such as, circulation fields, propagation of Rossby waves, water masses, front movement and the depth and temperature of mixed layer, and so on. At the present time, the most suitable and practical way to monitor the subsurface thermal conditions is through XBT measurements using the Voluntary Observing Ships (VOS). Actually, under this understanding, TOGA/WOCE XBT/XCTD Program Planning Committee (TWXXPPC) has negotiated arrangements to monitor the world oceans by VOS. Fig. 7 shows present TOGA/WOCE XBT monitoring network (WOCE IPO, 1992).

The northern North Pacific, that is, the PICES area has been monitored relatively very well since early 1970s. This monitoring (PX-26 in Fig. 7) has been done under the so-called "TRANSPAC" program operated by NOAA with the financial support of the US Navy from early 1970s.

By using the XBT data accumulated in this program, clarification of structure and its changes of the upper ocean has been made from various viewpoints. Among them, for example, Rossby wave propagation was observed in the North Pacific Current (White, 1982), the behaviour of the paths of the Kuroshio and the Kuroshio Extension were analyzed (Mizuno and White, 1983) and interannual variability in the heat content of the Kuroshio Extension associated with 1982 ENSO event was found (White and He, 1986).

However, since 1993 when the US Navy stopped funding to the TRANSPAC program, the number of XBT measurements is now decreasing: before 1992, approximately 3500 XBT profiles were available, while they were reduced to about 2000 in 1993 (see Fig. 8: NOAA NOS, 1994) and to about 1500 in 1994.

Therefore, as international or multinational program, the establishment of a new XBT monitoring program like as TRANSPAC is urgently required. TWXXPPC already recommended the establishment of TRANSPAC XBT pool like TOGA XBT pool (WOCE IPO, 1994).

The author believes that PICES can play an important role in the establishment of a new functional monitoring program in succession to TRANSPAC XBT program.

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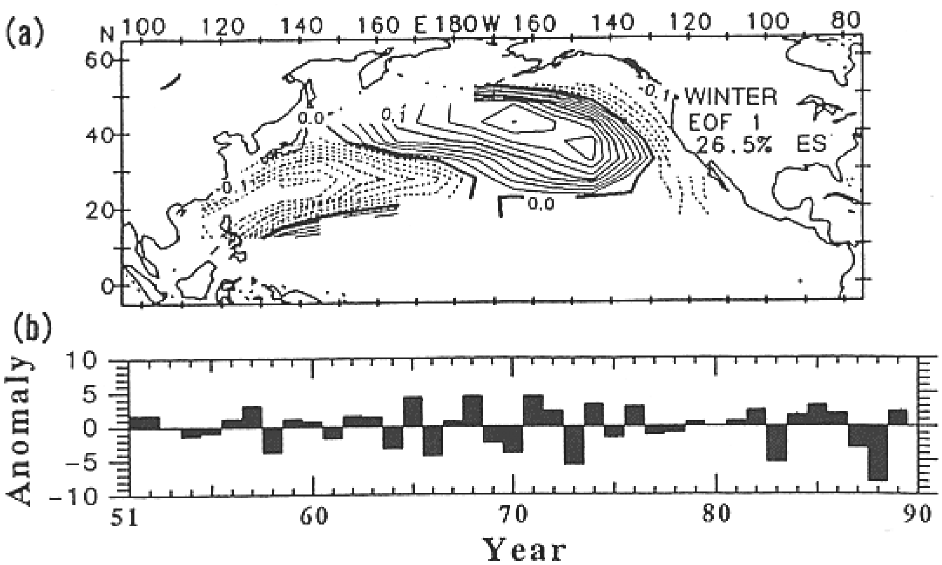

Fig. 1. Distribution map of the first EOF for bandpass (24-60 months: ENSO time scale) filtered winter SST anomalies (a) and its time coefficient (b). This mode can account for 26.5% for total variance of bandpass filtered winter SST anomalies. After Tanimoto et al. (1995a).

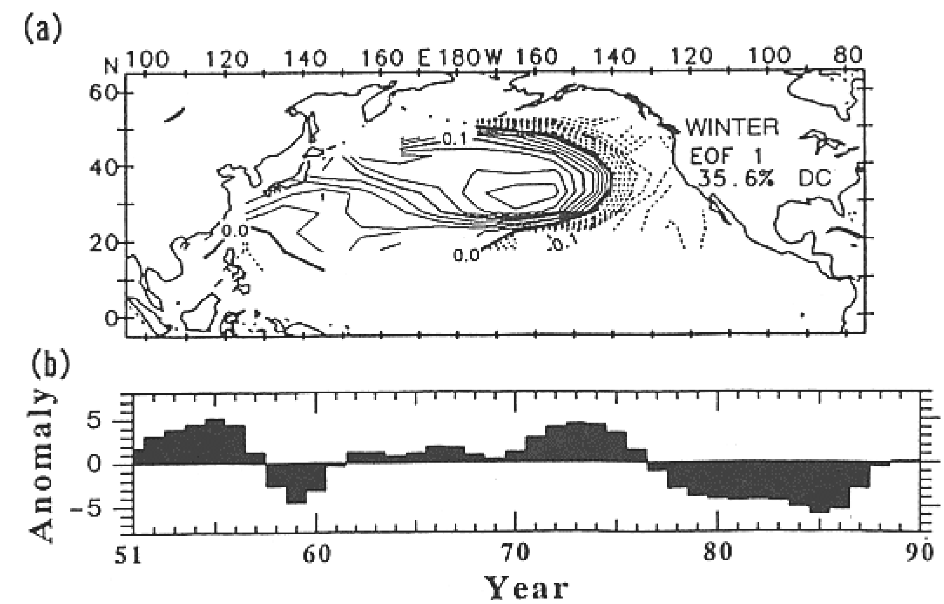

Fig. 2. Same as in Fig. 1 but for lowpass (60 months-longer: DC time scale) filtered winter SST anomalies (a) and its time coefficient (b). This mode can account for 35.6% for total variance of lowpass filtered winter SST anomalies. After Tanimoto et al. (1995a).

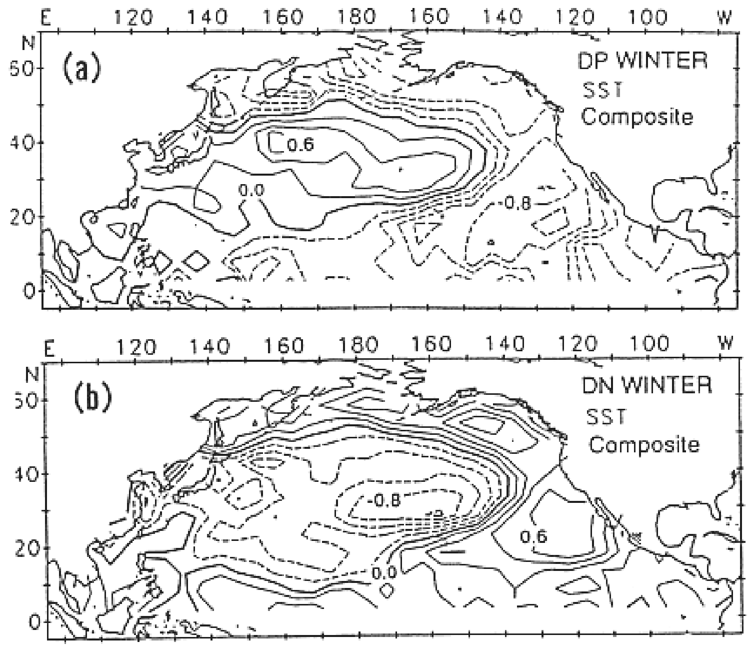

Fig. 3. Composite maps of unfiltered winter SST anomalies for the period of DP and DN. Here, the period of DP (DN) corresponds to the years with time coefficients greater (smaller) than one (minus one) standard deviation of the time coefficient of the first DC mode shown in Fig. 2(b). DP (DN) winters are winters of 1953-56 and 1972-75 (1959 and 1979-86). Contour interval is 0.2C. After Tanimoto et al. (1995a).

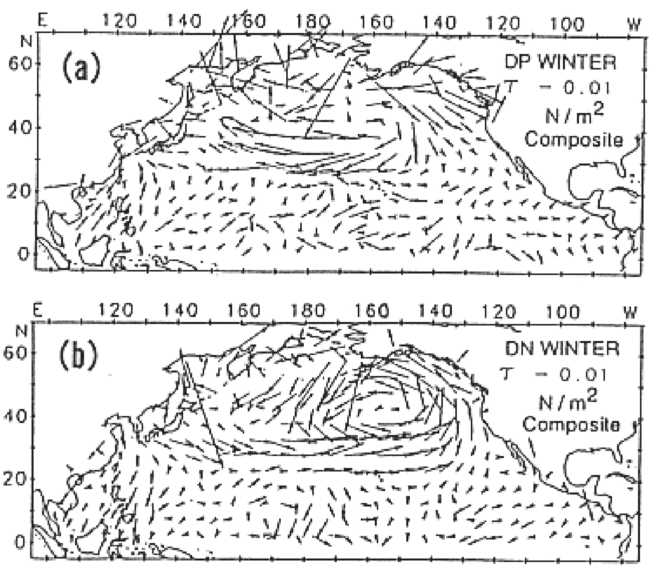

Fig. 4. Composite maps of winter (December through February) wind stress anomalies for the periods (a) DP and (b) DN. The length of arrow at upper right corner corresponds to the wind stress intensity of 0.01 Nm-2. After Tanimoto et al. (1995b).

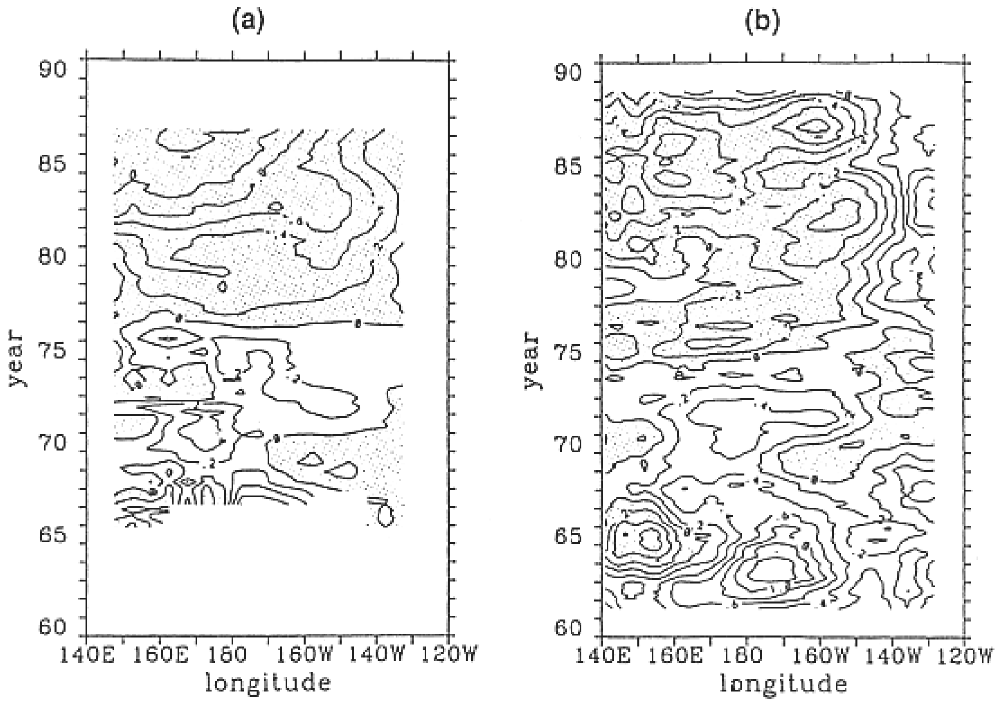

Fig. 5. (a) Time longitude diagram of 0-400m vertically averaged temperature anomalies along 35N-45N latitude band and (b) SST. Time series at each grid (5 x 5 degrees latitude x longitude box) are a 3-year running average. Contour interval is 0.2C and negative values are shaded. After Watanabe and Mizuno (1994).

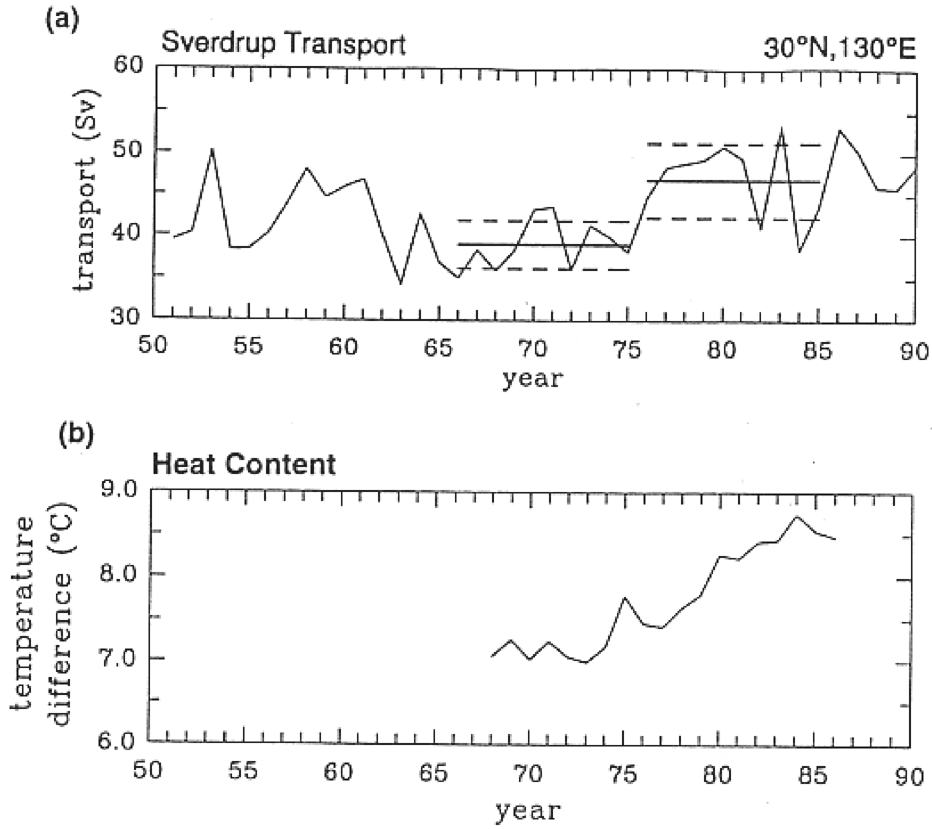

Fig. 6. (a) Time series of the annual mean Sverdrup transport at 30N, 130E. Means and the standard deviations for the two decades of 1966-75 and 1976-85 are indicated.

(b) Time series of the meridional difference of annual mean temperature vertically averaged from surface to 400m and zonally 150E-160E. The difference is taken for the values 29N minus 39N and smoothed with 3-year running mean. This time series can be regarded as that of heat content. After Yasuda and Hanawa (1995).

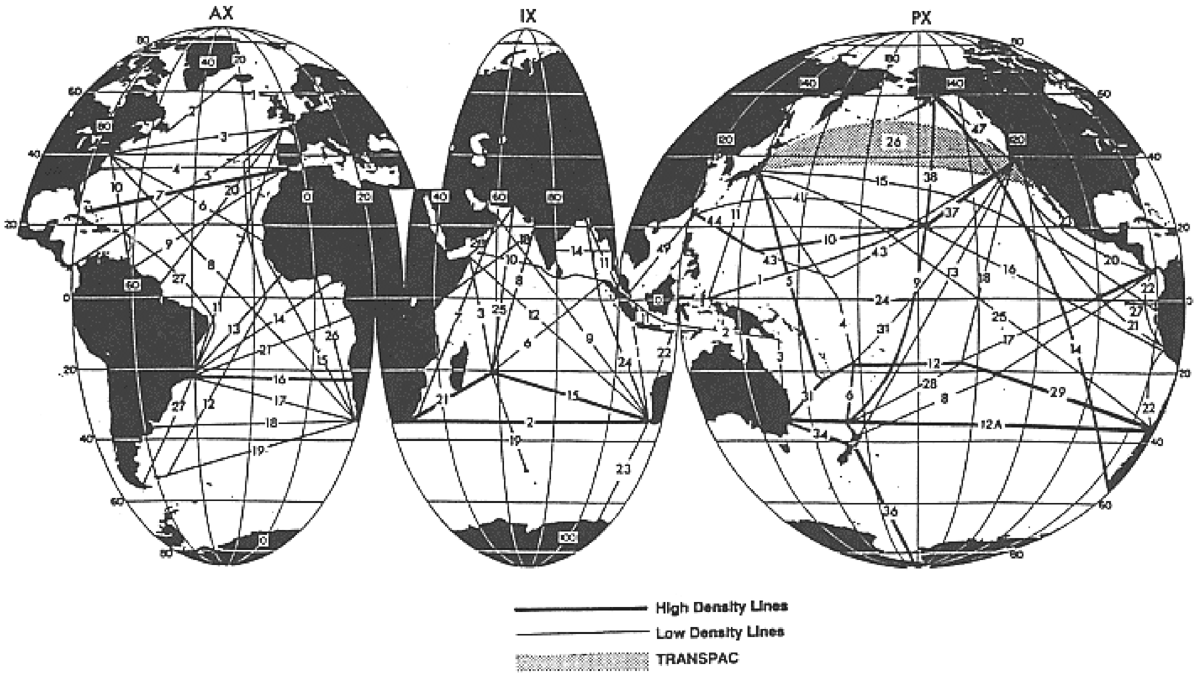

Fig. 7. TOGA/WOCE XBT monitoring network. PX-26 (envelope) corresponds to the TRANSPAC monitoring area. After WOCE IPO (1992).

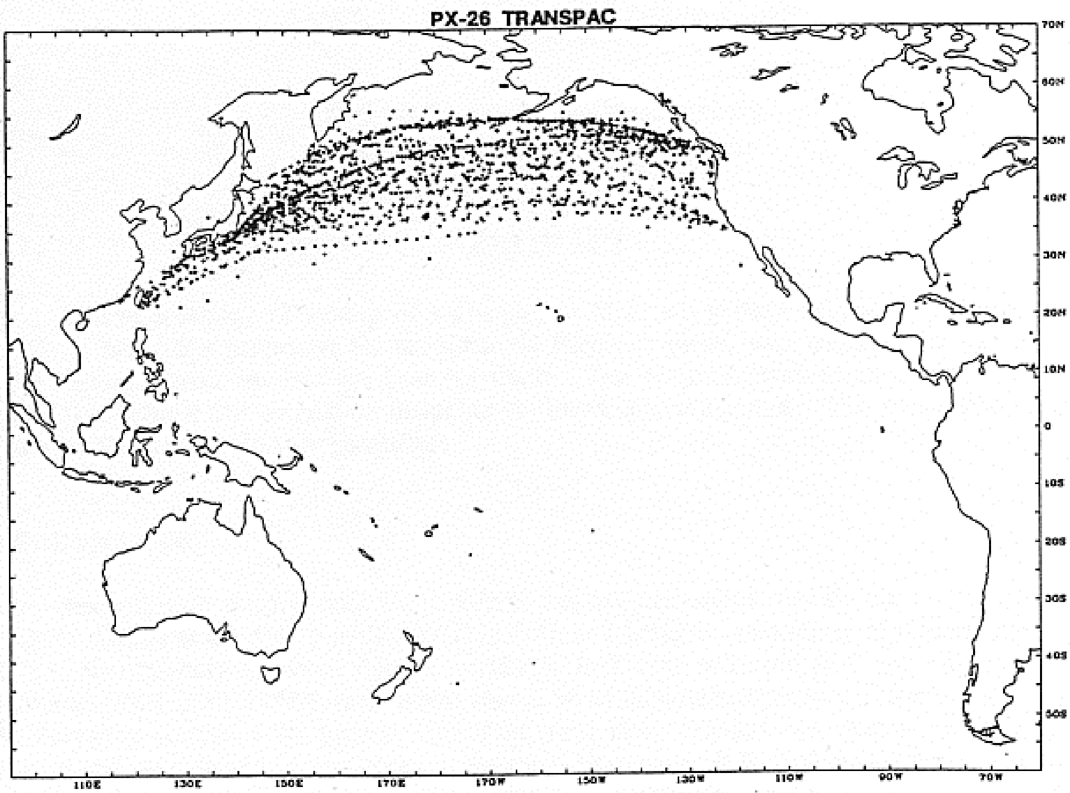

Fig. 8. Distribution of XBT observational points made in 1993 in the PX-26 (TRANSPAC) area. In this year, approximately 2000 XBTs were dropped.

Summary of Variability of Physical Conditions in the Subarctic North Pacific

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ABSTRACT

A brief summary of studies of variability in SST, sea surface height, atmospheric forcing, circulation, and water properties for the subarctic Pacific is given, with emphasis on the central and eastern regions. Most studies have been for the first few variables, with much less attention paid to salinity distributions and circulation including the barotropic component. The importance of salinity for near-surface stratification is quantified.

1. INTRODUCTION

Variability in the subarctic North Pacific has been reviewed recently for PICES (Denman et al., 1991; Okuda et al., 1991; PICES WG1, 1994), as a basis for discussions of climate change. The IOC has also recently reviewed decadal variations in Pacific subarctic SST and relations with the atmosphere (UNESCO, 1992); their report should be an important input to the monitoring discussion. This workshop continues the general discussion, with more emphasis on what is most likely to be unknown and what should be monitored in the subarctic region. It is presumed that the principal goal of this knowledge and monitoring is improvement in understanding, and perhaps even predictions, of interannual change in weather and fishery conditions in the subarctic region and surrounding land masses.

Much of the analysis of interannual and decadal variability in the North Pacific has been in terms of winds, sea surface temperature (SST), and major current transports. Although it is clearly recognized and stated that salinity is the dominant factor in the upper ocean stratification of the subarctic gyre, little emphasis has been given to its temporal variability. This is probably due to the meteorological and fishery bias of many of the analyses of the upper ocean, since SST is the principal item of oceanographic interest for atmospheric modeling and fishery predictions. A much smaller data base might also be a culprit, but there are data enough to at least begin such an analysis, and on which to base recommendations for future sampling.

The following paragraphs are intended to provide background for discussion of uncertainties and future monitoring, and are not intended to be an inclusive review of subarctic variability.

2. SURFACE VARIABILITY AND ATMOSPHERIC FORCING

The dominant mode of variability for the atmosphere and the ocean surface temperature in the subarctic North Pacific has been shown to be associated with the Pacific-North American (PNA) mode, which consists in the atmosphere of a strengthening of the Aleutian Low, centered near the dateline and 40°N (Davis, 1976; Horel and Wallace, 1981; Namias et al., 1988; Hanawa et al. 1989a,b; Hanawa, 1991; Trenberth, 1990). When the Aleutian Low is enhanced, the sea surface temperature is low in the western and central subarctic gyre, and high in the eastern Gulf of Alaska, presumably due to the enhanced cyclonic ocean circulation. More of the North Pacific Current flows northward into the Gulf of Alaska when the Aleutian Low is strong (Chelton and Davis, 1982). Hanawa (1991) has also shown that there is warming in the Kuroshio region as part of this dominant mode. SST in the subarctic gyre lags atmospheric pressure, unlike in the tropics where the ocean leads the atmosphere.

Despite the simplicity of the atmospheric PNA teleconnection pattern showing the downstream relation of the Aleutian Low to El Nino (Horel and Wallace, 1981), the strength of the Aleutian Low has been shown in recent years to be somewhat independent of El Nino. In particular there was a protracted period, from 1976 until 1988, in which the low was especially strong (Trenberth, 1990; Hanawa, 1991; Ebbesmeyer et al., 1991); this period included three El Nino's. Initiation of the shift in 1976 appears to have been due mainly to anomalous surface heat flux (Miller et al., 1994).

The southern limit of the subarctic circulation depends on the definition. Near the sea surface, the subarctic halocline, which provides the stratification between the mixed layer and the underlying ocean, outcrops within the subarctic frontal zone (Roden, 1991), which is 4° latitude wide on average (Yuan, 1994). The associated temperature front is an important salmon boundary. The subarctic frontal zone location has a several degree latitude range, but appears to have little systematic seasonal or interannual variation (Yoshida, 1993; Kazmin, 1994; Yuan, 1994). The salinity and salinity gradient associated with it also have little seasonal dependence (Yuan, 1994). The temperature gradient is stronger in summer (Kazmin, 1994). In all seasons the temperature and salinity gradients are nearly density-compensating across the front.

The subarctic front arises near the western boundary when the Oyashio separates from the coast. Even there it is a relatively density-compensated feature. The frontal zone is advected eastward across the Pacific, and slightly southward relative to the general circulation, probably due to Ekman advection to the right of the prevailing westerlies. In the eastern Pacific, the frontal zone lies several degrees south of the maximum average Ekman transport. It is generally characterized by two strong fronts, although in the western Pacific it sometimes appears as one, and in the eastern Pacific sometimes as three or more fronts. The fronts within the subarctic frontal zone may be maintained in their abruptness by the local Ekman transport, which is convergent through most of the year except the winter (and hence not in the annual average) (Yuan, 1994). The principal unknowns with respect to the subarctic front are what its true scale is (due to the lack of very high resolution measurements), whether it has a significant seasonal cycle (due to lack of winter and spring measurements), what maintains the abruptness of individual fronts within it if anything, why it is nearly density-compensating, and how the shallow salinity minima which extend to the southeast from the subarctic frontal zone are formed locally.

Below the surface layer, the southern limit of the subarctic region can be identified with the northern limit of the main subtropical salinity minimum, North Pacific Intermediate Water (NPIW). The boundary is coincident with the subarctic front in the western Pacific, but is several degrees north of it in the eastern Pacific (Talley et al., 1991; Zhang and Hanawa, 1993; Yuan, 1994), probably due to southward Ekman advection of the surface waters as stated before. Part of the NPIW may advect northward into the Gulf of Alaska (Musgrave et al., 1992; Talley et al., 1991; Zhang and Hanawa, 1991). Zhang and Hanawa (1991) looked at interannual variation of the boundary between the subarctic and subtropical water masses for 1978-1990, and found that the front was clear and identifiable from 170°E to the dateline, with variation of no more than 2°. There has been no systematic study of the temporal variability of this boundary in the eastern Pacific, and there has been no study of its seasonal variability anywhere.

The importance of salinity in the vertical and lateral property structure of the subarctic Pacific can be illustrated using the Turner angle (Ruddick, 1983) and an adaptation of it to lateral gradients of temperature and salinity. The Turner angle is $Tu=\arctan[(N_T-N_S)/(N_T+N_S)]$

where $N_T=\alpha \frac{\partial T}{\partial n}$ and $N_S=\beta \frac{\partial S}{\partial n}$ where n is either the vertical or lateral coordinate.

Fig. 1 shows the distribution of Turner angles for the surface layer, based on both the vertical stratification and the meridional gradients. It is clear that north of the subarctic front (about 40°N), both the vertical and meridional density structures are dominated by salinity. It appears essential then that any monitoring program must include salinity. It also would seem essential that any study of the seasonal, interannual, and decadal time scales in the subarctic region should include study of the salinity structure.

3. GULF OF ALASKA AND NORTHERN CALIFORNIA CURRENT

A short recent review of variability in the northeastern Pacific can be found in Kelly et al. (1993). In the eastern North Pacific the North Pacific Current bifurcates into northward flow around the cyclonic Alaskan Gyre, and southward flow into the California Current. The bifurcation point and strength of flow in the two directions is time dependent (Chelton and Davis, 1982). Interannual variation appears to be linked mainly to the strengthening of the Aleutian Low discussed above. The time scales are interdecadal, as indicated above, and there also appears to be a shorter period El Nino time scale, as documented by Van Scoy and Druffel (1993). They present evidence for flow of subpolar water into the California Current (to 30°N) during non-El Nino years (weak Aleutian Low), and no subpolar water at that location during El Nino years. They also indicate that vertical mixing in the Alaska gyre is more vigorous during El Nino years (strong Aleutian Low).

Royer (1993) suggests that about 30% of the interannual oceanic signal in the northeast Pacific is associated with the 18.6-year nodal tide. Substantiation is through comparison of recent ocean time series with air temperature, and then demonstration of the tidal signal in a 150-year record of air temperature. Because explanations for the non-El Nino interdecadal signal in the northern North Pacific (e.g. the 1976-1988 "regime") are so complex (Miller et al., 1994), it seems that the tidal ideas should receive consideration.

The winds and wind stress curl in the Gulf of Alaska have a strong seasonal cycle. However, the baroclinic transport of the Alaska Stream does not respond strongly to this seasonal cycle (Royer, 1981; Musgrave et al., 1992); instead it has a much stronger interannual variation (Royer, 1981). This is likely related to the turning latitude for the annual baroclinic Rossby wave, which lies south of the subarctic gyre. The fresh water runoff into the Gulf of Alaska, which contributes about 40% of the fresh water for the northeast Pacific, also has a strong seasonal cycle which is correlated with a seasonal cycle in baroclinic transport very close to the Alaska coast (Royer, 1982).

Barotropic response of the currents in the Gulf of Alaska to the winds, on the other hand, is well established for short time scales (e.g., Niiler and Koblinsky, 1985; Chave et al., 1992); no long time series of barotropic motion exist. Chave et al. show that the ocean's barotropic mode near 40°N in the central Pacific is coherent with several atmospheric variables; the response is non-local suggesting propagation of Rossby waves over the rough topography. Kelly et al. (1993) used the short existing Geosat altimetry record to show the two modes of the eastern Pacific: strong and weak California Current (Chelton and Davis, 1982), and that the variations in surface height are related to local Ekman pumping, rather than to Rossby wave propagation.

The only long time series which exists for the interior Gulf of Alaska is that at Ocean Weather Station Papa (50°N, 145°W) and the section east of it into the Canadian coast. Tabata et al., (1986) and White and Tabata (1987) describe the variability in steric height and deeper anomalies over the years, showing that heating controls the steric height offshore of Canada while the freshwater sources control it along the coast, and that there is westward propagation of anomalies with a period of 2-5 years, respectively.

Large eddies along the eastern boundary of the Gulf of Alaska have not received much attention. Tabata (1982) describes the eddy often found off Sitka. Another eddy is often found off Queen Charlotte. The NOARL model suggests a progression of semi-permanent large eddies along the eastern boundary. Their dynamical forcing and effect on the gyre circulation, if any, has yet to be considered.

4. WESTERN SUBARCTIC GYRE

Variability in the western subarctic gyre was reviewed by Okuda et al. (1992), and in the PICES Working Group 1 report (1994). The western subarctic gyre is fed from the east by the Alaska Stream, and may also contain an enclosed strong cyclonic circulation elongated along the Kuril Islands. The extent and time dependence of the connection with the eastern subarctic is not quantified, due to a lack of time series observations.

Until recently there appeared to be relatively few estimates of boundary current transports in this region other than around Japan. The WG1 report lists a set of recent baroclinic transport estimates for the East Kamchatka Current; the largest variations seem to be associated with the presence or absence of the large Kuril eddies. The East Kamchatka Current and Oyashio appear to have significant barotropic components, which have not been well measured at all. The Oyashio transport variations relative to standard depths are regularly documented, but do not include the barotropic component. Sekine (1988) shows that the strength of the southward intrusion of the Oyashio water into the Mixed Water Region has some relation to the variation in Sverdrup forcing over the subarctic Pacific, but variations in the baroclinic transport of the Oyashio are not easily related to Sverdrup forcing changes.

A series of large eddies (200 km diameter) along the Kuril Islands has been observed in various ways over the past decade (e.g., Bulatov and Lobanov, 1984; see PICES WG1 report). These eddies may propagate slowly northward, entraining subarctic waters. Their relation to the Oyashio and East Kamchatka Current has not been clarified, nor are their dynamics understood.

Variability in the circulation of the Okhotsk and Bering Seas has not been quantified, other than to indicate that there are changes. The sea ice cover in both marginal seas has been better described and related to interannual changes in atmospheric conditions. Overland et al. (1994) and Stabeno and Reed (1994) review and present what is currently known about the circulation of the Bering Sea, pointing out the considerable eddy activity within most of the region except along the boundaries. No time scales for variability are indicated. Likewise, it is apparent that the Okhotsk Sea has a great deal of eddy activity, which is likely to be responsible for mixing the various inputs such as the saline Soya Current water, inflowing North Pacific water, and ventilated shelf water.

5. SUMMARY

Large uncertainties regarding variability in the subarctic Pacific exist in a number of areas. A major one is the variability of salinity and the freshwater sources: the surface stratification is determined by salinity, with densest ventilation occurring where the waters are most saline (brine rejection due to sea ice in the Bering and Okhotsk Seas, and where saline subtropical waters are injected into the subarctic in the Okhotsk Sea and through Tsugaru Strait into the region adjacent to the Oyashio). Salinity probably plays a role in vertical mixing of surface properties downwards as well. It appears though that no systematic studies have been made, or might be possible with the available data set, of the correspondence of the circulation and SST with surface salinity. A second major area of uncertainty is the nature of the changes in the circulation, and the effect of the circulation on SST. The dominant atmospheric and SST modes have been established. There has been significant progress in looking at sea surface height. However, changes in the circulation, water properties including salinity, and barotropicity of subarctic currents (which appears to be large), as well as the role of the large eddies in the western and eastern boundary regions are much less well studied. Basic phenomena of the subarctic circulation and water properties are not understood: the formation and maintenance of the Kuril and eastern boundary eddies, the processes which create the subarctic front and maintain it across the Pacific, the maintenance of the nearly isothermal layer which extends through the halocline, quantification of the changes in circulation and ventilation due to sea ice formation and melt, and so on.

Future monitoring efforts must concentrate on the interannual changes of the subarctic ocean as they relate to atmospheric variables. The barotropic response must be included, which thus involves much shorter time scales as well. The principal processes controlling the strength of the Aleutian Low should be elucidated, including the complicated relationship with El Nino, which has a much shorter time scale. Salinity cannot be ignored in any such studies. Modeling efforts, including sufficiently realistic topography because of the importance of the barotropic ocean response and a realistic inclusion of salinity as an independent variable must continue to be encouraged. Mixing of temperature and salinity must be properly parameterized and modeled. Models synthesizing altimetry and in situ measurements should be developed as the model physics become better honed. Observations must be continued, both for providing basic information for understanding processes which are not now understood, as well as for providing the basic long time series needed for monitoring and improving models.

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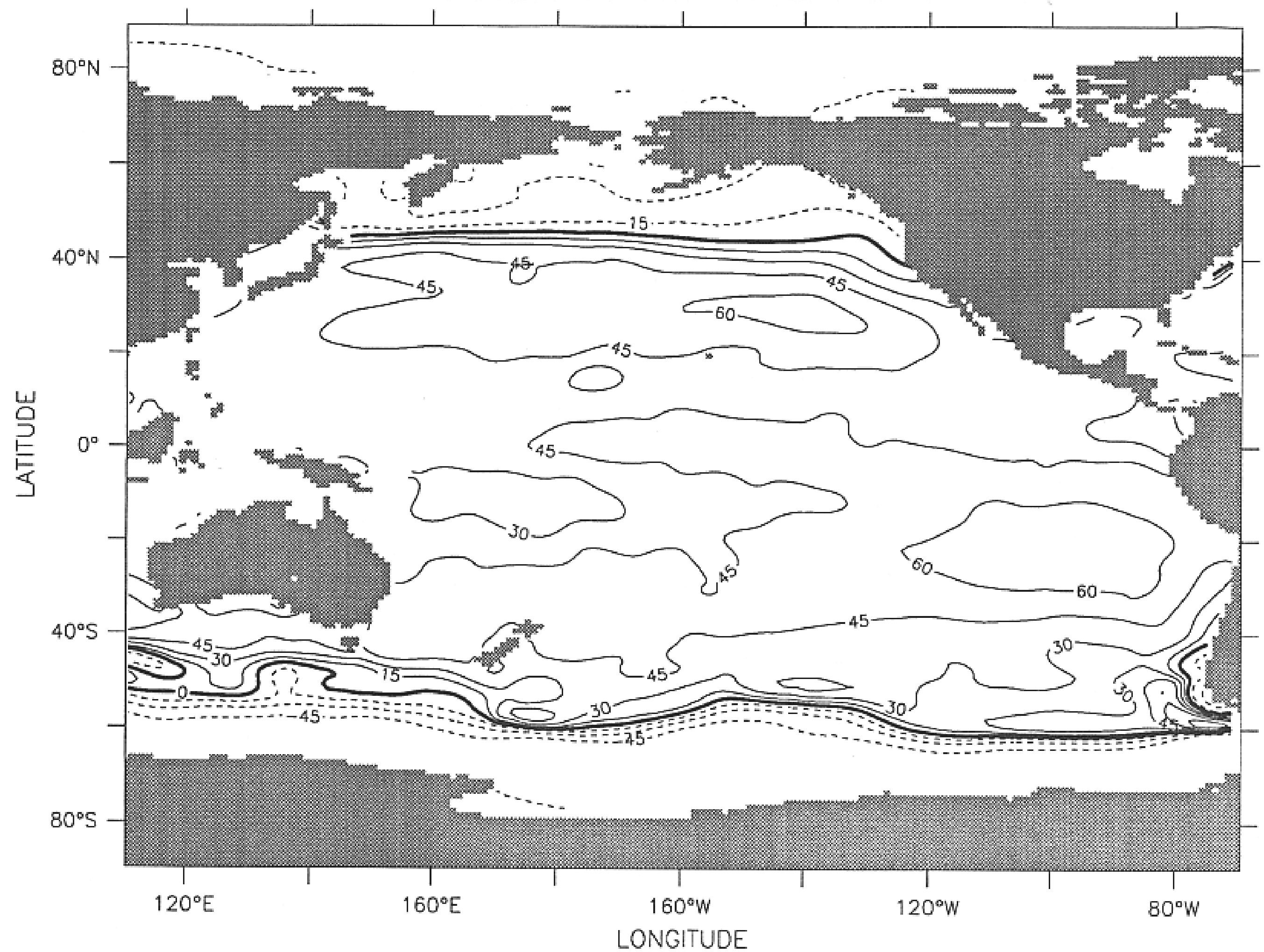

Turner angle (10m over 200m) annual average

Fig. 1.(a) Turner angle based on vertical differences of temperature and salinity, calculated as a simple difference of the 200 meter values from the 10 meter values, from the annually- average Levitus (1982) data. Angles between 45 and 90 degrees indicate the predominance of salinity in the vertical density gradient. Angles less than -45 degrees occur when there is a temperature inversion.

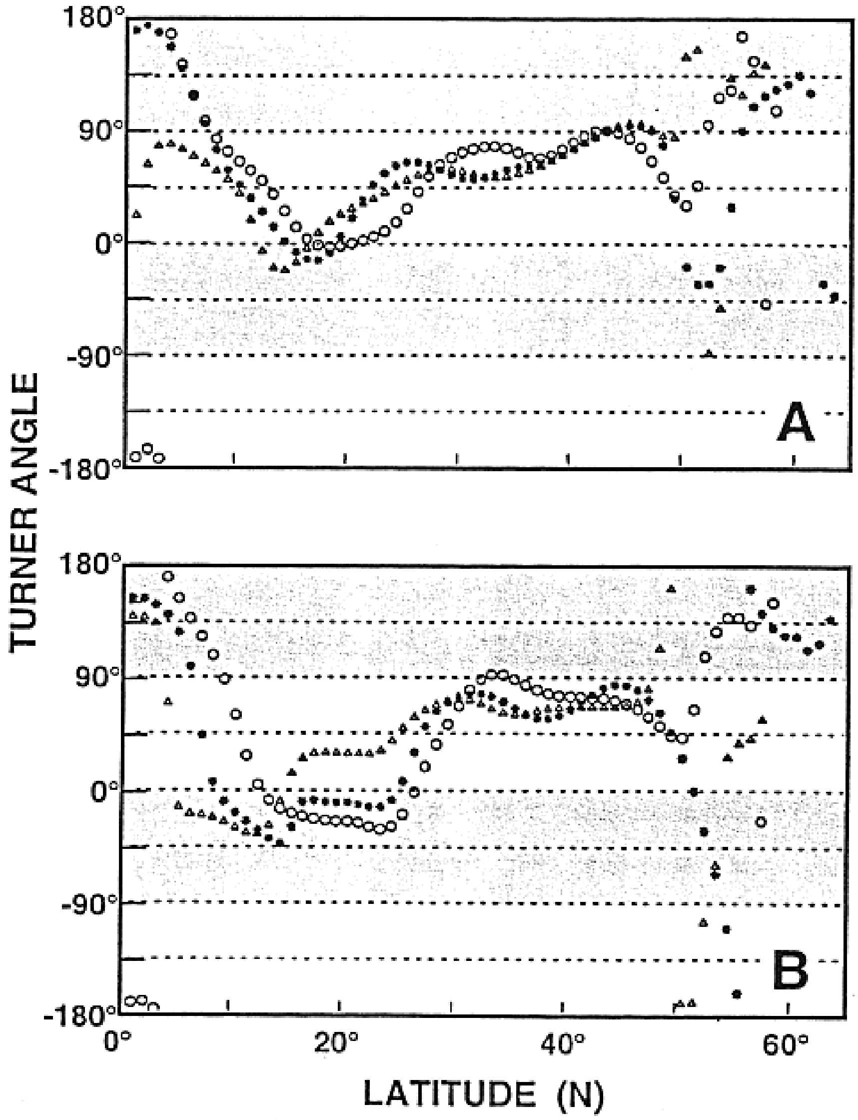

Fig. 1.(b) Turner angle based on meridional variation of temperature and salinity at 10 m depth in the winter and summer, also from the Levitus (1982) annually-averaged data. Turner angles greater than 90 degrees and less than -90 degrees indicate salinity dominance - in the shaded regions. The different symbols denote western (triangles), central (asterisks) and eastern (circles) North Pacific. (Yuan, 1994.) (The Turner angles for the vertical and meridional calculations used different conventions for the angles.) For both (a) and (b) the crossover from temperature to salinity dominance occurs around 45 N, coinciding with the subarctic frontal zone.

Seasonal and Interannual Variability of Ice Cover in the North Pacific Marginal Seas

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Due to thermal inertia, ice conditions of the Okhotsk, Bering and Japan Seas are one of the most expressive indicators for variability of the climatic system in the whole North Pacific. One of the main generalized characteristics of ice conditions is ice cover (Maksimov, 1977; Plotnikov, 1987; L. Gidrometeoizkdat, 1977).

Most essential input to the evolution of ice processes in these frozen seas is made by seasonal and interannual fluctuations (Plotnikov, 1987; Zubakin, 1987). Still, for these seas, the quantitative analyses of similar variability needs further study.

As the initial data, mean 10-day values of the ice cover in the Pacific marginal seas were used starting from 1960. These values were obtained by averaging maps of ice surveys from aircraft performed in the region, and then they were corrected with the help of all data available (vessel observations, satellite data, etc.) The time interval of studies included the period from December to May for the Okhotsk and Bering Seas, and from December to April for the Japan Sea. The choice of such periods is determined by climatic peculiarities of the considered water areas (L. Gidrometeoizdat, 1977).

1. SEASONAL VARIABILITY

Seasonal fluctuations of the ice cover conditions in the Pacific marginal seas are mainly conditioned by astronomical forcing and possess a vividly expressed annual period. These oscillations are superimposed by fluctuations conditioned by of regional hydrometeorological factors.

To analyze the processes of the seasonal evolution for the ice cover, a theory of periodically correlated random processes was used (Zakharov, 1981). To represent the seasonal character of ice processes the typical curves were computed of seasonal motion of ice cover for mild, normal and severe winters were drawn (Fig. 1), with an estimation of the standard deviations (Fig. 2) that characterize their variability.

Distribution of estimates shows a vividly expressed annual motion of the ice cover and a local minimum of variability during the periods of maximal development of ice conditions. Variability maxima are generally related to the periods of development and destruction of the ice cover except the Bering Sea, where variability maximum appears to be connected with the start of the period of the ice destruction. Differences in the analyzed characteristics are expressed first in absolute values and in terms when different phases start to take place.

Thus, the ice cover maximum in the Okhotsk Sea takes place in March, that of the Japan Sea in February; and for the Bering Sea, there are, two periods of possible maxima of the ice cover: at the end of February and early of April.

Some similarity of the ice processes in the Okhotsk Sea and the Japan Sea, as well as distinct differences in the Bering Sea (in statistical sense) are probably related to the nature of the evolution of the large-scale system of hydrometeorological parameters of the North Pacific (L. Gidrometeoizdat, 1977).

To estimate seasonal variability, the correlation matrices of the ice cover distribution interrelation were constructed (Table 1-3). These matrices show that the correlations during the ice season generally preserve their sign. This fact reflects the stability (in a statistical sense) of tendencies arising in the ice processes. The maximal sluggishness of the processes (the largest duration of significant relations satisfying 95% level of significance) is related to the periods when ice cover is a maximum. For the Okhotsk Sea this period lasts from February to March; and for the Bering Sea, from the end of January to April; and for the Sea of Japan, from January to April. The duration of significant correlations during these periods reaches 3-4 months (i.e. state of ice conditions just during these periods influences further evolution of ice processes). The least sluggishness of the processes occurs at the beginning of the ice period. This can be easily explained by the initial nonstability of the ice processes at the beginning of ice formation, when even small disturbances of the external factors can lead to radical redistribution of the new formed ice cover. To a lesser extent, such a fact is also a characteristic for the end of the ice period .

2. INTERANNUAL VARIABILITY

To estimate long-term variability, the many years ice cover series for the same 10-day periods were prepared. They were used to calculate statistical estimates of the correlation functions and frequency spectra. They permit us to determine to a first approximation some peculiarities of the long-term variability, as well as to repeat the main energy-carrying frequencies and their amplitudes.

The schemes show (Fig. 3) that the time of relaxation of the autocorrelation functions generally exceeds the data time spacings (one year). Along with this, the autocorrelation functions show the presence of the quasiperiodic components (Fig. 4). The periods of 2-3 years, 7-8 years, 11 years and 22 years are distinguished. With some confidence it is possible to distinguish the 2-3 year variability, considering the data available (33 years).

Investigations performed recently (the INPOC Programme, etc.) permit us to suggest that such periodicity is possibly typical of the whole climatic system in the Pacific Ocean Subarctic Zone. A possible interpretation of the mechanism for formation of such variability, at least as a first approximation, can be presented as a chain of cause and effect relationships.

The disturbance of the thermodynamic equilibrium in the Subarctic Front zone in September of 1991 lead to the increase of the interlatitudinal heat- and mass-exchange, which in turn increased the level of inhomogeneity of the typical subarctic water and lead to sharp contrasts of temperature between the ocean-atmosphere and the continents in winter of 1991 - 1992. (This conclusion is made on the basis of analyzing the cruise data obtained in the bounds of the INPOC Programme and it is presented in the Report on 60 Cruise of RV "PRILIV", 1993).

Along with the noted characteristics, the whole complex of the hydrometeorological parameters, including heat storage of the upper 1,000 m layer of the ocean, was changed. The next fact is intensification of circulation activity and enforcing of heat-exchange processes between the atmosphere and the ocean, which lead to the extreme cooling of the active layer in the North Pacific in winter of

1991-1992. Secondary effect of the stated processes was expansion of the area of the western subarctic subtype of waters in spring of 1992 (Fig. 5). (Many researchers distinguish the western and the eastern subtypes of the subarctic waters. The main difference between these subtypes consists of the fact that the western subtype possess a distinct subsurface cold layer, and the eastern one does not possess it, or it is very weak. Corresponding to these subtypes, the subarctic zone of the Pacific Ocean can be divided into two areas - western and eastern. Here the location of the division can be considered as some indicator of the hydrological processes in the Subarctic zone of the Northern Pacific). Until the fall of 1992 the cold intermediate layer formed as a result of winter convection, was preserved, and the area of its distribution even increased to some extent due to advection and mixing processes. Thus, for the winter of 1992-1993 the western area had lower heat storage of waters and the active circulation in the system of the western marginal currents transporting the cold water. This lead to decreasing of temperature contrasts between the ocean, atmosphere, and the continents. The given changes caused a decrease of heat exchange at the sea surface and a decrease of dynamic activity, which lead to relatively mild winter conditions. Correspondingly, the margin of the western subtype of waters in winter of 1993 shifted to the west as compared to that a year before.

In the eastern sector, according to the proposed suggestions, the processes were more diverse. This is indirectly show by the results of the successive comparison of smoothed temperature values on the isopicnic surfaces along 50° N, as well as values of their difference from one survey to another. This gives the opportunity to suggest the presence of some balance mechanism between the thermohaline characteristics of the western and eastern zones, separating the subarctic zone of the North Pacific into two parts.

In view of such notions it is possible to explain the phase opposition of the ice processes in the Japan and Okhotsk Seas compared to the Bering Sea, as well as its exceptions. This is related to the position of the boundary between the western and the eastern subtypes of waters of the subarctic zone. The phase opposition is explained by the location of the mentioned seas in the zones of predominance of different subtypes of the subarctic structure of waters. A possible shift of the boundary between the waters subtypes far to the east, when the predominant part of the Bering Sea would be located in the area of the western subtype of waters, would result in a shift in ice processes in all North Pacific marginal Seas.

The peaks of spectral density occurring at the periods of 11 years and 22 years, probably are related to the heliophysical factors. This periodicity is marked in many climatic systems also.

Considering the problem of classification by the variability character, we can distinguish as a separate group the initial period of ice formation including December, where the correlation and especially spectral functions differ considerably from the same functions during the other periods. This is also the reflection of the non-stability of ice processes during the initial period of ice formation. Consequently, the ice cover during the period of its primary formation cannot be considered as indicator of the climatic processes. In space and time, a distinct non-regularity of the weight distribution of some cyclic components of variability is observed. While during the periods of the ice growth the process of 7-8 year periodicity prevail, during the ice destruction periods it is observed the shift of energy-carrying characteristics to the area of lower frequencies and the fluctuations of 11 years and 22 years periods start to predominate. It is noted the more expressed predominance of 7-8 years variability in the Okhotsk and Japan Seas and 11 and 22 years variability in the Bering Sea.

Besides the noted periodicities there is a series of secondary cycles, which probably are overtones of main oscillations, and their relative input to their total variability is considerably lower. The large-scale variability of ice cover in the Pacific marginal seas is a useful guide, to possible changes in climatic conditions, and similar variations are observed by researchers for the other hydrometeorological characteristics as well.

Taken together it is hoped that such studies can make useful additions to our understanding and of the functioning of a complicated climatic system as the North Pacific.

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Table 1. Correlation matrix of seasonal ice interconnections (Okhotsk Sea)

	10	December			January			February			March			April			May		
Month	day	I	II	III	I	II	III	I	II	III	I	II	III	I	II	III	I	II	III
December	I	1.00																	
	II	0.70	1.00																
	III	0.38	0.67	1.00															
January	I	0.24	0.48	0.79	1.00														
	II	0.21	0.32	0.65	0.79	1.00													
	III	0.11	0.38	0.49	0.67	0.73	1.00												
February	I	0.31	0.41	0.41	0.59	0.56	0.88	1.00											
	II	0.40	0.44	0.43	0.56	0.65	0.89	0.95	1.00										
	III	0.41	0.43	0.48	0.50	0.64	0.84	0.87	0.94	1.00									
March	I	0.32	0.29	0.31	0.38	0.38	0.67	0.79	0.79	0.88	1.00								
	II	0.29	0.33	0.29	0.33	0.31	0.65	0.75	0.70	0.82	0.91	1.00							
	III	0.20	0.25	0.26	0.37	0.46	0.80	0.78	0.81	0.90	0.88	0.88	1.00						
April	I	0.23	0.31	0.34	0.34	0.54	0.79	0.75	0.81	0.87	0.80	0.77	0.91	1.00					
	II	0.16	0.26	0.42	0.37	0.52	0.78	0.76	0.79	0.83	0.72	0.69	0.80	0.85	1.00				
	III	0.09	0.10	0.27	0.24	0.44	0.68	0.64	0.66	0.72	0.66	0.61	0.70	0.74	0.91	1.00			
May	I	0.04	0.04	0.27	0.29	0.40	0.56	0.48	0.47	0.58	0.58	0.60	0.61	0.61	0.72	0.85	1.00		
	II	0.18	0.25	0.37	0.44	0.40	0.61	0.50	0.49	0.60	0.50	0.62	0.60	0.61	0.65	0.70	0.87	1.00	
	III	0.51	0.33	0.23	0.37	0.43	0.30	0.33	0.39	0.46	0.34	0.42	0.34	0.32	0.20	0.21	0.38	0.61	1.00

Table 2. Correlation matrix of seasonal ice interconnections (Bering Sea)

	10	December			January			February			March			April			May		
Month	day	I	II	III	I	II	III	I	II	III	I	II	III	I	II	III	I	II	III
December	I	1.00																	
	II	0.83	1.00																
	III	0.49	0.65	1.00															
January	I	0.16	0.40	0.75	1.00														
	II	0.25	0.55	0.68	0.83	1.00													
	III	0.25	0.43	0.59	0.54	0.78	1.00												
February	I	0.06	0.27	0.53	0.49	0.61	0.79	1.00											
	II	0.29	0.36	0.39	0.27	0.40	0.52	0.71	1.00										
	III	0.26	0.39	0.40	0.84	0.45	0.55	0.59	0.79	1.00									
March	I	0.07	0.20	0.19	0.19	0.36	0.39	0.32	0.65	0.73	1.00								
	II	0.01	0.14	0.20	0.28	0.42	0.42	0.41	0.71	0.66	0.85	1.00							
	III	0.21	0.40	0.50	0.37	0.56	0.65	0.61	0.75	0.60	0.71	0.84	1.00						
April	I	0.22	0.39	0.57	0.44	0.51	0.61	0.66	0.78	0.60	0.65	0.73	0.90	1.00					
	II	0.26	0.43	0.57	0.35	0.46	0.57	0.59	0.76	0.68	0.70	0.70	0.88	0.91	1.00				
	III	0.21	0.37	0.54	0.31	0.43	0.51	0.63	0.75	0.67	0.68	0.67	0.83	0.84	0.90	1.00			
May	I	0.20	0.39	0.40	0.40	0.58	0.63	0.60	0.70	0.69	0.73	0.72	0.85	0.80	0.88	0.86	1.00		
	II	0.12	0.32	0.34	0.39	0.51	0.55	0.51	0.70	0.65	0.77	0.73	0.79	0.83	0.84	0.77	0.92	1.00	
	III	0.00	0.19	0.16	0.22	0.26	0.34	052	0.67	0.51	0.52	0.56	0.56	0.69	0.71	0.62	0.71	0.77	1.00

Table 3. Correlation matrix of seasonal ice interconnections (Japan Sea)

Month	10 day	December I	December II	December III	January I	January II	January III	February I	February II	February III	March I	March II	March III	April I	April II	April III
December	I	1.00														
	II	0.59	1.00													
	III	0.22	0.49	1.00												
January	I	0.03	0.29	0.59	1.00											
	II	0.16	0.42	0.30	0.64	1.00										
	III	0.02	0.39	0.56	0.57	0.76	1.00									
February	I	0.64	0.21	0.31	0.57	0.62	0.73	1.00								
	II	0.23	0.35	0.28	0.46	0.71	0.71	0.73	1.00							
	III	0.12	0.08	0.01	0.26	0.69	0.50	0.38	0.66	1.00						
March	I	0.07	0.00	0.13	0.09	0.51	0.51	0.47	0.53	0.75	1.00					
	II	0.12	0.04	0.11	0.12	0.52	0.42	0.38	0.56	0.81	0.77	1.00				
	III	0.37	0.30	0.27	0.18	0.57	0.59	0.38	0.64	0.74	0.72	0.66	1.00			
April	I	0.41	0.18	0.12	0.06	0.49	0.46	0.41	0.36	0.46	0.57	0.37	0.61	1.00		
	II	0.40	0.32	0.31	0.09	0.44	0.56	0.33	0.61	0.47	0.33	0.40	0.67	0.47	1.00	
	III	0.66	0.36	0.34	0.15	0.22	0.19	0.25	0.38	0.12	0.02	0.17	0.43	0.44	0.53	1.00

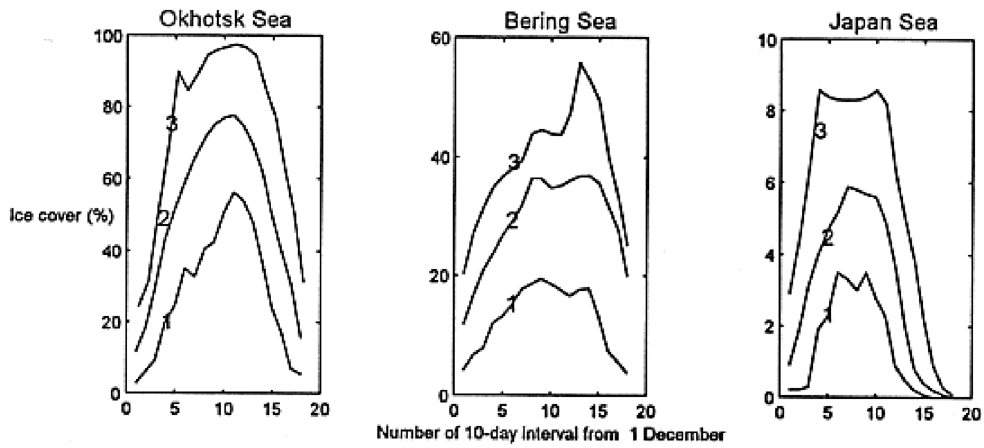

Fig. 1. Seasonal distribution of the extremal small (1), normal (2) and extremal large (3) ice cover.

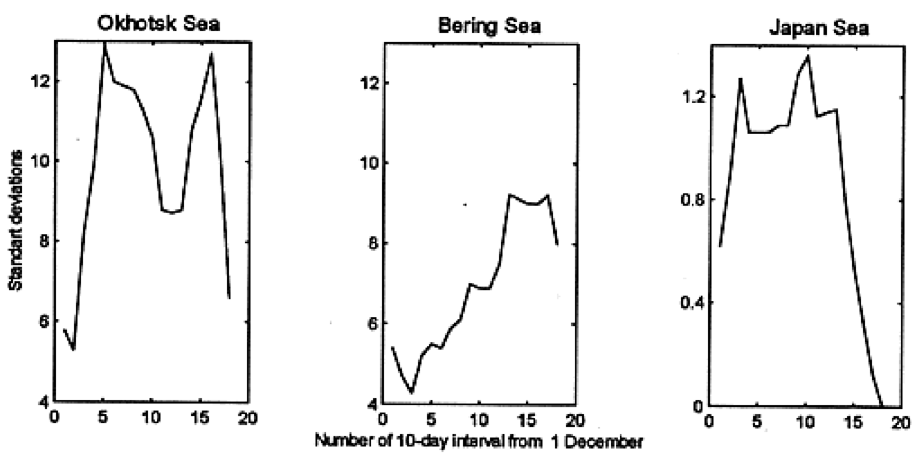

Fig. 2. Seasonal standard deviations of the ice cover distribution.

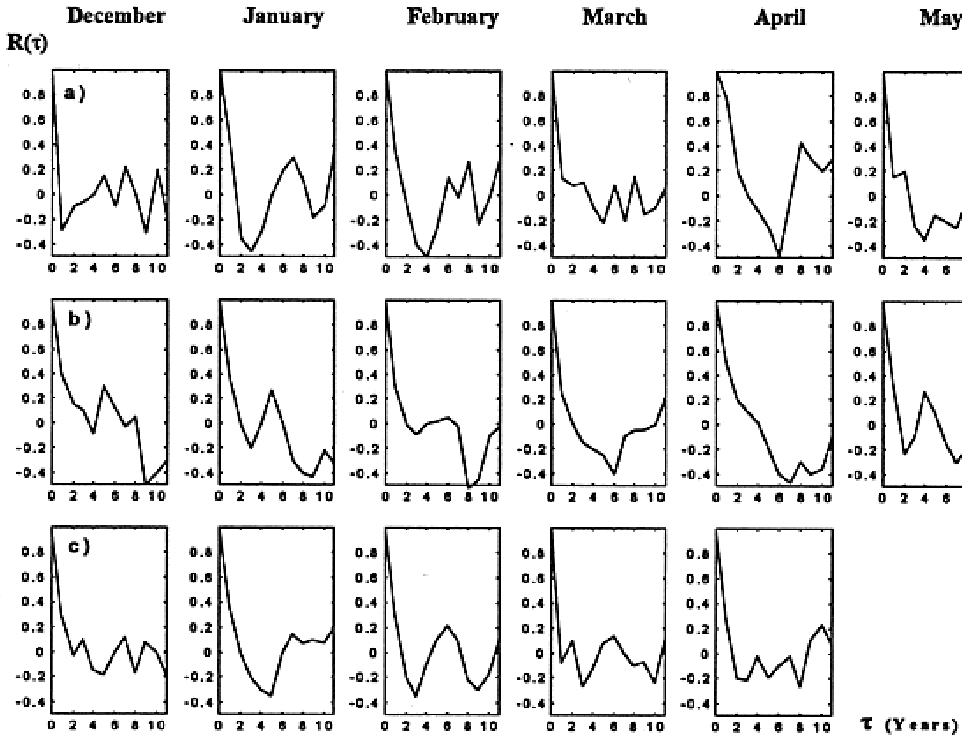

Fig. 3. Estimates of ice cover correlation function for Okhotsk (a), Bering (b) and Japan (c) seas.

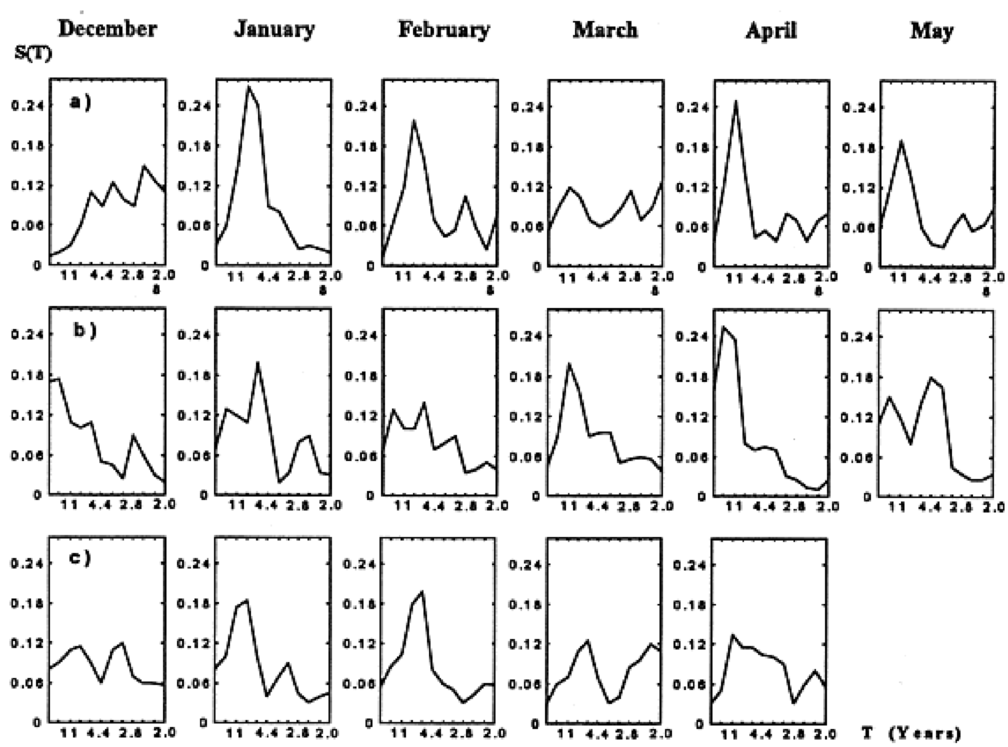

Fig. 4. Estimates of ice cover frequency spectra for Okhotsk (a), Bering (b) and Japan (c) seas.

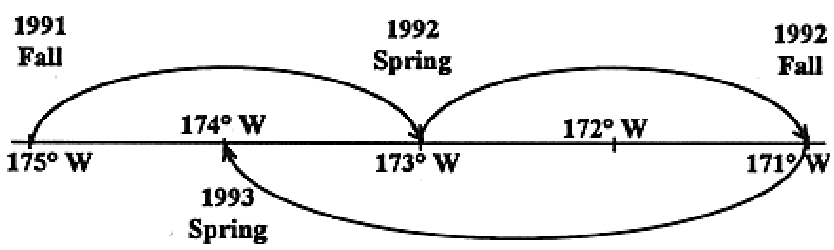

Fig.5. Location of the boundary between western and eastern subarctic water subtypes on 50° N during 1991-1993.

Space and Time Scales of Variability in the Subarctic North Pacific: Implications to Monitoring the System

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1. INTRODUCTION

In recent years a great deal of new scientific exploration of the North Pacific Ocean and its marginal seas has taken place. This exploration has yielded a number of new observations that will undoubtedly change our basic understanding of the general circulation of the Pacific and its biological and chemical consequences. On the other hand, many of these new observations are one-time measurements, though addressing many of the most basic scientific questions concerning the North Pacific circulation implies a knowledge of the *change in the circulation over time* and the spatial variations in these changes; to collect the measurements required to answer such questions necessitates a dedicated program to monitor the Pacific circulation over a relatively long period of time at many locations. The design of such an observing system and its ultimate implementation will require a major international committment to planning and resources. This paper will attempt to elicudate some of the more elementary sampling considerations that must be taken into account in the initial design of a monitoring program, in the context of several contemporary examples of variability of the North Pacific subarctic circulation.

2. SAMPLING CONSIDERATIONS

Defining a sampling strategy requires a knowledge of the particular scientific questions to be posed. What variables, at which sites, are desirable to monitor? Some specific examples of variables having special scientific importance, from several locations in the subarctic North Pacific, will be given in later sections of this paper. Initially, however, it is necessary to examine the degree to which useful quantitative estimates of relevant quantitites can be made. Beyond purely descriptive analyses, there are a number of rather standard quantitative estimates that can be used in measurement interpretation.

2.1. *Estimating the mean and variance*

The simplest quantity that can be used to characterize a random variable is its mean value. Suppose that a normally-distributed oceanographic random variable ϕ is measured as a function of time, t. In the discussion in this section ϕ will be considered to be a scalar quantity such as temperature or sea level; for a vector variable such as velocity each component may be treated separately as a scalar by the methodology given here. The variable ϕ is assumed to be a continuous function of time, $\phi(t)$, that is sampled at N various discrete times t_i so that $\phi_i = \phi(t_i)$. Then the estimated mean value of ϕ, $\tilde{\phi}$, is simply

$$\tilde{\phi} = \frac{1}{N} \sum_{i=1}^{N} \phi_i \quad . \tag{1}$$

The difference between the true mean value of ϕ, $\bar{\phi}$, and the estimated mean $\tilde{\phi}$ is the standard error in the estimate of the mean of ϕ, ϵ_{ϕ}, and is given by the well-known result

$$\epsilon_{\phi} = \sigma_{\phi} \sqrt{\frac{2I_{\phi}}{N}} , \tag{2}$$

where σ_{ϕ} is the standard deviation of the variable ϕ and I_{ϕ} is the *integral time scale* of the autocorrelation function of ϕ; it is a measure of the number of samples of ϕ over which the variable becomes uncorrelated with its previous state. If $I_{\phi} = 1$, then each sample is independent of those that came before it, and the measured series ϕ_i contains N independent samples. In general, however, this will not be the case, as often the sampling rate is designed to be considerably faster than the decorrelation rate.

The autocorrelation function of ϕ, R_{ϕ}, can be defined as

$$R_{\phi}(\tau) = \frac{1}{\sigma_{\phi}^2 T} \int_0^T \phi(t)\,\phi(t+\tau)\,dt , \tag{3}$$

where T is the duration in time of the variable ϕ and τ is the lag time. For short lags ($\tau \sim 0$), $R_{\phi}(\tau) \sim 1$ (ie, the correlation is very high); for long lags ($\tau \rightarrow \infty$) the correlation decays to zero. The integral time scale I_{ϕ} is a measure of the decorrelation time averaged over the entire series and is given by

$$I_{\phi} = \int_0^{\lambda} R_{\phi}(\tau)\,d\tau , \tag{4}$$

where λ is the longest lag in the correlation function. The integral time scale is thus a measure of the net total area under the autocorrelation function. I_{ϕ} is generally not readily discernable from a visual inspection of ϕ; it is not just the typical time scale of variability of ϕ and must generally be computed according to (4) rather than simply estimated from a cursory examination of the data.

Using these definitions, it is possible to estimate the quantity of data required to discern the mean value of ϕ to within acceptable error bounds. Suppose, for example that is desired to estimate the annual mean sea surface temperature θ at at point in the subarctic Northeast Pacific to an accuracy of .2°C. The standard deviation of temperature about the mean over the annual cycle is about 2°C (see the sea surface temperature maps in Halpern et. al., 1994). If we take I_{θ} to be 20 days (a typical value for the western North Atlantic, where a number of such studies have been carried out), then the number of days of data N required for the desired level of accuracy is given by

$$N = 2I_{\theta}\frac{\sigma_{\phi}^2}{\epsilon_{\phi}^2} = (2)(20)\frac{(4)}{(.04)} = 4000 . \tag{5}$$

Thus, assuming that the sea surface temperature was normally-distributed and statistically stationary, samples would have to be collected at daily intervals over a duration of 4000 days (nearly 11 years) in order to estimate the mean temperature at a single point to the required accuracy. Of course, the temporal duration of this monitoring experiment could be decreased if a spatially averaged sea surface temperature was an acceptable measure of the temperature at a point. If an ensemble average is substituted for a time average in (1), then an *areal average* could incorporate data from many points in some region, as long as the samples were all independent in both space and time. One could in principle define a *spatial* integral scale analogous to (4) that would help to prescribe the spatial separation between samples necessary for independence, although determining the spatial correlation function analogous to (3) would in practice be difficult. Since the required temporal duration of an experiment utilizing sampling in time at a single point seems generally prohibitive [from (5)], it seems likely that some mixture of space and time sampling, resulting in mixed areal/temporal averages of relevant quantites, will be necessary in any large-scale oceanic monitoring effort.

Placing quantitative bounds on the accuracy of the mean estimate requires, from (2), an estimate of the variance of ϕ. If ϕ is normally-distributed, then the variance is given by

$$\sigma_\phi^2 = \frac{1}{T} \int_0^T [\phi(t) - \bar{\phi}]^2 \, dt \ . \tag{6}$$

In an analogy with (2), the error in estimating the variance, $\epsilon_{\mathrm{var}\,\phi}$, can be shown to be

$$\epsilon_{\mathrm{var}\,\phi} = \sigma_\phi^2 \sqrt{\frac{2J_\phi}{N}} \ , \tag{7}$$

where the quantity J_ϕ is the *square-integral time scale* and is given by

$$J_\phi = \int_0^\lambda R_\phi^2(\tau) \, d\tau \ . \tag{8}$$

Due to the nature of the integrals in (4) and (8), it can be shown that generally $J_\phi > I_\phi$. This implies through (2) and (7) that the relative error in estimating the variance of ϕ will usually exceed the relative error in estimating the mean value. This can be generalized to estimates of higher moments of ϕ as well: the higher the moment, the more data that will be required to estimate that moment to within acceptable error. In addition, errors in estimating the variance will feed back into errors in estimating the mean value, since through (2) and (7) and error in estimating the variance of ϕ will lead to an error in determining the standard error of the mean.

2.2. *Estimating the spectrum of variability*

It is often useful to characterize a data record by examining its variance as a function of frequency, or its *spectrum.* Many standard references on spectral analysis exist; a useful example is Bendat and Piersol (1971). To review briefly, suppose that the Fourier transform of $\alpha(t)$ (a function having zero mean) is denoted as $\hat{\alpha}(\omega)$, where ω is the frequency. The Fourier transform is defined as

$$\hat{\alpha}(\omega) = \int_{-\infty}^{\infty} \alpha(t) \, e^{-i\omega t} \, dt \ . \tag{9}$$

The ***autospectrum*** of $\alpha(t)$, $\Phi_{\alpha\alpha}(\omega)$ (usually known as simply the spectrum) is then

$$\Phi_{\alpha\alpha}(\omega) = \left|\hat{\alpha}^2(\omega)\right| = \hat{\alpha}(\omega)\,\hat{\alpha}^*(\omega) \ , \tag{10}$$

where the asterisk denotes the complex conjugate. Equation (9) is written for the case where $\alpha(t)$ is a continuous function in time; in actuality, all oceanographic variables are sampled over some finite duration T, at regular intervals t_i separated in time by an interval Δt. In any finite data record there will be N total samples, so that $N\Delta t = T$. The result is that the discretely sampled version of (9), rather than being continuous in frequency space, would instead yield samples at the frequencies 0, $1/T$, $2/T$, ..., $(N/2)/T$, for a total of $N/2+1$ samples in frequency space. The highest frequency, known as the *Nyquist* frequency ω_{Ny}, can be written as

$$\omega_{Ny} = \frac{N}{2T} = \frac{N}{2N\Delta t} = \frac{1}{2\Delta t} \ . \tag{11}$$

This is the well-known *Sampling Theorem*: for a discretly sampled record, the lowest frequency that can be resolved (other than zero) is determined by the record length T; the resolution in frequency space (ie, the distance between adjacent frequencies) is $1/T$; the highest frequency that can be resolved is determined by the sampling interval Δt.

In practice, individual spectral estimates in (10) are usually averaged in some way (ie, either by averaging over several frequency bands or by averaging spectra from several different pieces of the record together) in order to reduce the noise in any individual estimate. Since $\alpha(t)$ is sampled discretely over a finite duration, the spectrum $\Phi_{\alpha\alpha}(\omega)$ is only an *estimate* of the true spectrum of α. The error in estimating the spectrum ϵ_Φ (ie, the rms difference between the true spectrum and the measured spectrum) is approximately [see Bendat and Piersol (1971) for a complete derivation]

$$\epsilon_\Phi(\omega) \approx \Phi_{\alpha\alpha}(\omega)\left(\frac{1}{T\Delta\omega} + \frac{\Delta\omega^4}{576} \left[\frac{1}{\Phi_{\alpha\alpha}(\omega)} \frac{\partial^2\Phi_{\alpha\alpha}(\omega)}{\partial\omega}\right]^2 \right)^{1/2} , \tag{12}$$

where $\Delta\omega$ is the distance between spectral estimates after some filtering and averaging in frequency space has been applied. This result suggests that errors in spectral estimation are controlled by two competing effects. The first term in the brackets on the right side of (12) represents the estimation error due to random fluctuations in the spectrum and is inversely proportional to the bandwidth of each spectral estimate; increasing the bandwidth will presumably increase the degree of smoothing of the spectrum and correspondingly decrease the estimation error. The second term on the right in (12), negligible for white or linearly varying spectra, becomes important near sharp peaks in the spectrum; this term increases proportionaly with the square of the bandwidth and quantitatively shows the effect of too little resolution in the vicinity of spectral peaks.

As an example of the expected errors, consider the spectrum of sea level at San Francisco, based on a data record nearly 100 years long, as shown by Wunsch (1981; his figure 11b). The annual peak in this spectrum has a spectral amplitude of approximately 3×10^4 cm^2/(cycle per day). The spectrum has been smoothed so that the bandwidth is approximately 1 (month)$^{-1}$. With a record 100 years long, the *first term* of the error ϵ_Φ from (12) can be estimated as

$$\varepsilon_\Phi \approx (3 \times 10^4\,\mathrm{cm}^2/\mathrm{cpd})\left(\frac{8.3 \times 10^{-2}\,\mathrm{yr}}{(100\,\mathrm{yr})}\right)^{1/2} \approx 864 \ \mathrm{cm}^2/\mathrm{cpd} \ . \tag{13}$$

For most monitoring applications, this would be a small and acceptable error (about 2%; even including the second term in (12) would only increase ϵ_Φ in this case by a factor of about 3) and demonstrates the utility of monitoring the ocean at a single point over a very long time when it is feasible.

An important consideration in any discussion of spectral analysis is the question of the choice of the sampling interval Δt, which determines the Nyquist frequency, ω_{Ny}. In theory, ω_{Ny} is chosen by choosing

Δt so that little or no variance lies at frequencies above ω_{Ny}. If there is energy in the true spectrum at frequencies above the Nyquist frequency, this energy will appear spuriously at frequencies *below* ω_{Ny} due to the well-known phenomenon known as *aliasing* (see Bendat and Piersoll [1971], Chapter 7, for example, for a complete discussion). If the energy at frequencies above ω_{Ny} is small, then this is not a significant problem. If the energy is large, however, as might be the case if there were a significant peak in the spectrum at some frequency $\omega > \omega_{Ny}$, aliasing can potentially be a serious problem. Consider the situation, for example, with the semidiurnal tide (M_2), having a period of 12.42 hours (ω = 1.93 cycles per day). There is a peak in the spectrum at M_2 in nearly all sea level, surface and subsurface velocity, and subsurface temperature records from the world ocean. It seems likely that most efforts to monitor the ocean over long space and time scales will probably sample the ocean at intervals Δt too large (> 6.2 hours) to yield $\omega_{Ny} > \omega$ in this case. Thus, the M_2 peak, if it exists for a particular sampled variable, will be aliased into the estimated spectrum at lower frequencies. For a sampling interval of 1 day, the M_2 peak will appear at a period of 14.8 days in the estimated spectrum; for a sampling interval of 10 days, the M_2 peak will be aliased to a period of 31 days (alarmingly near to a period of one month, which is the averaging interval used in many contemporary studies of ocean climate variability); and for a sampling interval of 30 days the M_2 peak will be aliased to a period of 1035 days, or 2.8 years. This suggests that care needs to be taken when designing a monitoring progam that potential tidal aliasing problems do not contaminate the data collected. For satellite altimeter studies, surely one of the cornerstones of any large-scale montitoring program, the M_2 tide can often be adequately removed by using a tidal model and may not be a serious problem. Similarly, there is not usually a strong tidal signal in mid-ocean sea surface temperature records, so it is possible that aliasing will not be a problem in this case either. For studies involving measurements of velocity however, or for any efforts to monitor the *subsurface* ocean, where there are often sizable M_2 signals arising from internal gravity waves, tidal aliasing is a potentially serious problem that should be investigated in more detail as part of the design of any monitoring scheme.

2.3. *Optimal mapping of oceanographically relevant fields*

The goal in many studies of large-scale ocean circulation will be to monitor the state of the ocean well enough to produce maps of various quantities of interest such as sea surface temperature or sea level. If the environment can be measured at a number of independent points in space, then in principle these data can be used to produce objective maps of the measured fields. The technique of *optimal interpolation*, or *objective analysis* is well-known in many branches of geophysics. A good example of its use in the mapping of oceanographic fields is given by Bretherton, Davis, and Fandry (1976).

Suppose that it is desired to use this technique to produce a mapped field of the oceanographic variable $\eta(\mathbf{x},t)$ at some time t_o, where $\mathbf{x} = (x,y)$ (ie, horizontal east and north coordinates). Further, suppose that the field η is sampled at N locations $\mathbf{x}_i$ in the presence of random noise. It is assumed that η has a mean value of zero. With these constraints in mind, it is possible to show that $\hat{\eta}$, the optimal, least squares estimate for the field η at a given time, and ϵ_m, the spatially dependent root-mean-square mapping error in the estimate of η, are

$$\hat{\eta}(\mathbf{x}, t_o) = \sum_{i,j} \langle \eta(\mathbf{x}, t_o)\, \eta(\mathbf{x}_j, t_o)\rangle \left[\langle \eta(\mathbf{x}_i, t_o)\, \eta(\mathbf{x}_j, t_o)\rangle\right]^{-1} \eta(\mathbf{x}_i, t_o) \tag{14}$$

and

$$\epsilon_m(\mathbf{x}, t_o) = \left\{ A^2 - \sum_{i,j} \langle \eta(\mathbf{x}, t_o)\, \eta(\mathbf{x}_i, t_o)\rangle \langle \eta(\mathbf{x}, t_o)\, \eta(\mathbf{x}_j, t_o)\rangle \left[\langle \eta(\mathbf{x}_i, t_o)\, \eta(\mathbf{x}_j, t_o)\rangle\right]^{-1} \right\}^{1/2} \tag{15}$$

where $A^2 = \langle \eta^2(\mathbf{x}, t_o) \rangle$. Here the brackets denote the spatial average of the quantity inside, and the superscript -1 denotes the inverse of the appropriate matrix. If the statistics of η are not time-dependent, then the rms error in mapping the field, ϵ_m, is also independent of time. If the spatial covariance matrix of the field, $F(\mathbf{r})$, defined as

$$F(\mathbf{r}) = \langle \eta(\mathbf{x}, t)\, \eta(\mathbf{x} + \mathbf{r}, t) \rangle \ , \tag{16}$$

is assumed to be independent of time, then all of the bracketed quantites in the definitions of $\hat{\eta}$ and ϵ_m are independent of time as well. Thus, the rms error in estimating η at any point on the map is only a function of the spatial covariance of the field and the locations of the observations, $\mathbf{x}_i$. This property can be used to estimate the mapping error of any variable whose covariance matrix is known, *even in advance* of carrying out a measurement program, and thus is potentially a boon to the design of a program designed to monitor the ocean.

There are many examples of the use of objective analysis in the historical literature; an relevant example that shows the errors in a hypothetical experiment that might be used to monitor temperature in the upper ocean over the North Pacific from ships of opportunity is given by White and Bernstein (1979). As shown in figure 1 (reproduced from the paper), the ratio of the mean-square mapping error to the variance of upper ocean temperature is a maximum where sampling is the least dense and decreases to acceptable values (<50%) as the sampling is decreased to 50 km intervals in the western Pacific and 200 km intervals in the eastern Pacific, the difference being mainly due to the large eddy signal in the west. White and Bernstein conclude that sampling at higher density than this in either region of the North Pacific is not cost effective, as the ratio of the error to the signal decreases only very slightly for denser spatial sampling. Of course, these results assume that the temperature data are collected at nearly the same time (synoptically) across the entire North Pacific; the temporal correlation function for large-scale North Pacific temperature given by White and Bernstein shows a first zero crossing near 10 months at all depths from 0–300 m and suggests that the temperature field should be monitored in such a way that each map contains data from a time interval much shorter than 10 months. This does not seem to be a serious limitation in the case of ships (there are usually a number of ships crossing the N. Pacific between 30°N and 50°N at any given time, and the crossing requires less than two weeks) and is certainly not a limitation if the sampling were to be carried out from orbiting platforms.

3. CONTEMPORARY SAMPLING ISSUES

The analysis techniques and observational parameters described above can be used to examine a wide range of scientific problems from an ocean monitoring perspective. But which problems are of the most immediate oceanographic interest, and which research programs are most likely to result in significant new knowledge? Answering these questions will require considerable debate in the international oceanographic community, by both the scientific personnel most likely to be involved and also at the administrative and governmental level in many different nations. In this work, it is more appropriate to examine some of the recent, intriguing observations of the North Pacific that may be used to suggest a sampling strategy for future monitoring of the Pacific circulation. This examination will especially focus on sampling the variability of the subarctic North Pacific from seasonal time scales (ie, a few months) to decades, in both the ocean interior and in boundary regimes.

3.1. *The North Pacific interior: the upper ocean*

While extensive observations on the low-frequency variability (ie, periods greater than a few days) of the mid-latitude N. Pacific do not generally exist, it is possible to qualitatively describe the interior circulation. Consider the hypothetical frequency-wavenumber spectrum of upper ocean variability shown in figure 2 (taken from Stewart et. al., 1986). This figure was originally produced during the planning stages of the Topex/Poseidon altimetry mission in order to attempt to predict the types and strength of sea level variability that might be sensed by an orbiting satellite and is based on a synthesis of a number of regional, mesoscale studies of mid-ocean variability carried out in the N. Atlantic in the 1970s and early 1980s. While in general similar information is not yet available for the Pacific, it is reasonable to assume that a frequency-wavenumber spectrum for the mid-latitude Pacific might share many of the qualitative features of figure 2.

It is suggested from figure 2 that the frequency-wavenumber spectrum of ocean variability is generally *red* (ie, increasing in the direction of the lowest spatial and temporal frequencies of variability), with relatively little variability at spatial scales less than the internal Rossby radius of deformation (a length scale dependent upon the stratification, water depth, and rotation of the Earth; approximately 15–30 km in the Subarctic N. Pacific) at time scales less than about 10 days. The spectrum of variability increases from these short space-time scales to a local peak in the spectrum in the near 100 days and 100 km. This variability of the ocean near this spectral peak represents the contributions of *mesoscale eddies*. This type of variability has been extensively studied in the N. Atlantic (less so in the Pacific), and there now exists a reasonably well-developed theoretical understanding of this phenomenon [see Rhines (1976) for a good review]. Figure 2 suggests that the broad peak due to eddies is one of the two most energetic features of the entire frequency-wavenumber spectrum. The other energetic feature is an energetic "line" in the spectrum at the annual period, extending from spatial scales of 100 km to the width of the basin. According to the prototype spectrum in figure 2, variability of the ocean *decreases* at very long temporal periods and spatial scales. However, there are to date few observations that can be used to test this idea, and, indeed, there is a growing appreciation that a great deal of energy (such as ENSO-related variability) probably exists at very low frequencies that cannot presently be well-resolved in spectral terms.

Which of the interior ocean phenomena depicted in the spectrum in figure 2 are candidates for long-term monitoring, and what techniques and strategies might be used to sample these phenomena? It is useful to briefly examine each of the major features in the spectrum shown in figure 2 in terms of the sampling methodology that might be required to monitor these phenomena.

3.1.a. *Mesoscale eddies*

Regional measurements of mesoscale eddies have been made at numerous sites in the world ocean, extending from the sea surface to the seafloor. Carrying out such measurement programs has inevitably required a major committment of equipment, ships, and technical personnel, even to measure the eddy field at a small number of fixed sites, and only rarely have such measurements been continued over durations longer than two years. It has been possible in recent years to measure the mesoscale variability at the sea surface from spatial scales of a few kilometers out to the basin scale using the technique of satellite altimetry, as shown recently by Wunsch and Stammer (1994). It is clear from their study that there is a relatively strong, seasonally modulated eddy field at the surface of the subarctic N. Pacific. The eddies generally exist at scales of a few hundred kilometers and, in the subarctic gyre of the N. Pacific, are especially strong in the Kuroshio Extension region and along the Aleutian arc. It is projected that in the near future such maps will be made at intervals of a few weeks, with accuracies of 2–3 cm. Thus, it appears that the eddy field at the sea surface can presently be monitored reasonably well, in the sense that the major mesoscale features at the sea surface can be located and followed over the entire subpolar N. Pacific gyre. If desired, objective maps of these fields can be constructed at intervals of a few weeks, to within acceptable error (by equations (14) and (15)), and, by averaging results over a long period of time, highly accurate estimates of the eddy variance and the spectral peak can be made (by equations (6), (7), and (8)). The quality of the variance estimates, and our knowledge of long-term modulation of the mesoscale eddy field, is limited only by the fact that the Topex/Poseidon mission has a design life of 5 years, suggesting that the satellite should continue to operate through 1997. It is obviously important that a Topex/Poseidon follow-on mission be carried out in order to continue to be able to monitor the sea surface of the world ocean in this manner.

Monitoring other mesoscale eddy-related parameters at the sea surface, such as sea surface temperature (SST) or velocity, will be considerably more difficult. It is possible to measure SST from satellites by the Advanced Very High Resolution Radiometer (AVHRR) technique; however, there is considerable error inherent in this method ($\sim .5°C$), leading to low-quality objective maps [from equation (15)], and the method is limited to cloud-free conditions. For the subarctic N. Pacific the latter restriction is especially important: it is estimated that less than 30% of the area of subpolar gyre is cloud-free on an annually averaged basis, with the cloud cover approaching 100% in winter months. As Niiler (1994) and Reynolds (1994) have recently impressively shown, both sea surface temperature and velocity can be monitored at daily intervals from an array of satellite-tracked surface drifters over an entire ocean. However, the number of drifters required to monitor the subpolar gyre of the N. Pacific at eddy-scale resolution is quite large. As an example, the surface area of the N. Pacific from 30°–50°N and from 135°E–135°W is approximately 7×10^6 km^2. In order to resolve variations at horizontal length scales greater than 100 km in just the subpolar gyre, approximately 700 surface drifters would be required, a number perhaps beyond the limits of managability. If the requirement for resolution were relaxed to 500 km, the number of surface drifters required would decrease to about 28 drifters, a more managable number, but with considerably less spatial resolution. In actuality, to account for possible instrument malfunctions and dispersion out of the subarctic N. Pacific, approximatly twice this number, or 60 drifters, might be necessary. Such an array might be useful for monitoring the SST and velocity fields on the largest spatial scales (> 1000 km), both in mapping the instantaneous fields and collecting long term statistics, but would be of little use for mapping the mesoscale fields. Eddy statistics (ie, velocity and SST variance) could be estimated from such an array, although it might be necessary to support such an array for many years in order to reduce the error inherent in the variance estimate to acceptable values. Using equations (7) and (8), and taking J_ϕ to be 20 days (a value based on N. Atlantic experience), suggests that such an experiment would have to continue for 1000 days in order to reduce the relative error in the variance map to 20%, assuming that the variance was stationary in time. For a seasonally varying eddy variance signal, a considerably longer experiment would be required. It is clear, moreover, that *is feasible* with present technology to monitor the surface properties of the subarctic N. Pacific at mesoscale resolution, although such a monitoring effort might be prohibitively expensive and unnecessary.

3.1.b. *Low-frequency variability*

At periods longer than a few months and scales longer than a few hundred kilometers, it is hypothesisized that the spectrum of variability decreases before increasing again near the region of the annual cycle, as suggested by figure 2. Theory would suggest that at these relatively low energy levels and long spatial and temporal scales that the variability in the ocean interior might be Rossby waves, although there is little observational basis to confirm this notion. The barotropic Rossby wave dispersion relation for a flat bottom ocean is

$$\omega = \frac{-\beta k}{k^2 + l^2} , \tag{17}$$

where ω is the wave frequency, k and l are the east and north wavenumbers, and β is the planetary beta effect, equal to $2\Omega \cos\theta / R_E$, where Ω is the rotation rate of the Earth and R_E is Earth's radius. This can be used to estimate the period of the highest frequency wave barotropic wave that can exist in the subpolar N. Pacific; if we set $l = 0$, so that only east-west waves are present, and take $k = 2\pi/L_P$, where L_P is the scale width of the subarctic gyre, then it is found that the highest possible frequency ω_{max} for a barotropic, planetary Rossby wave in the subarctic N. Pacific is

$$\omega_{\max} = \frac{-\beta L_{\mathrm{P}}}{2\pi} \quad . \tag{18}$$

Taking $\beta = 2\times10^{-3}$ km^{-1}day^{-1} and $L_P \approx 7000$ km yields a value for ω_{max} of 2.2 day^{-1}, or a period of about *14 days*. By a similar calculation, it is possible to show that the shortest period that a *baroclinic* Rossby wave can have in the N. Pacific poleward of 30°N is around *5 years*. To the extent that the low-frequency variability in the subarctic N. Pacific is governed by planetary Rossby wave-type dynamics, these periods place useful constraints on the types of low-frequency variability that might be expected.

3.1.b.1. *The annual period*

As shown in Halpern et. al. (1994), the seasonal cycle in SST over the N. Pacific can be seen to be a gradual increasing of temperature towards the North in late winter, leading to a maximum at the highest latitudes in late summer, followed by a large-scale cooling that again yields minimum values. This heating and cooling tends to proceed in a generally meridional fashion, with little evidence of any zonal asymmetry, leading to the conclusion that the mechanism of heating at the annual frequency is generally direct heat storage and release by the ocean rather than a complicated dynamical adjustment process. This heating appears to occur on a large-scale, and can be sampled by AVHRR techniques (pixel size is a few kiolmeters) with an expected error of approximately .5°C where the cloud cover is minimal. In general, however, the subarctic N. Pacific is covered by clouds, except in September and October, so that the AVHRR technique may not be of great use through most of the year. As with the eddy fields, sea surface temperature can in principle be well-measured from ships of opportunity (White and Bernstein, 1979) and objectively mapped to the sufficient accuracy to carry out useful studies of the annual cycle. However, in the winter months it is possible that the number of these ships falls below the number required for achieving, say, error levels of 50% in the maps. Accordingly, it once again appears that the use of an array of satellite-tracked surface drifters equipped with SST-measuring capability is the most cost-effective and reliable method for discerning SST over the subpolar N. Pacific gyre. As noted above, an array of approximately 60 such drifters should be sufficient to monitor the interior of the subarctic N. Pacific at approximately 500 km resolution. These drifters could also be equipped with meteorological packages so that humidity, surface air pressure and temperature, and wind speed could also be measured, aiding in the estimation of some of the more important ocean-atmosphere fluxes. Monitoring of this type is already taking place in the N. Pacific as part of the WOCE and TOGA programs, and it is hoped that is can be continued into the future with the requesite number of instruments in the subpolar N. Pacific.

The other important variability worth examining at the annual period in the ocean interior is sea level. This can be measured using satellite altimetry as with Topex/Poseidon as long as *absolute* sea level is not required (due to uncertainties in the shape of the geoid). Monitoring *fluctuations* in sea level in the the large-scale interior is being carried out with more-than-acceptable accuracy the present time, at approximately 10 day intervals, and will continue as long the Topex/Poseidon satellite continues to operate. It is also possible to monitor sea level using island-based tide gauges, and this is generally being done in the N. Pacific as part of the WOCE program. However, there are few mid-ocean sites suitable for these gauges in the interior of the subarctic N. Pacific, and it appears that this technique will not generally be useful as a tool, although it will be of great use in boundary regions and marginal seas where many gauges presently exist.

3.1.b.2. *Interannual variability*

At even lower temporal frequencies the monitoring problems are more difficult, simply because the observations must be continued for a longer time, although the *spatial* resolution need not be any greater than for monitoring higher frequency variability. It is just such long time series of measurements that are required in order to examine the nature of climate change in the physical environment and its resulting effects on oceanic ecosystems. It is not yet known *how long* such measurements must continue, although it is known that there is climate variability (and presumably oceanic variability as well) at the lowest frequencies imaginable. The prototype sea level spectrum shown in figure 2 suggests that at very low frequencies (periods > 1 year) the variability in the ocean decreases, but it is now clear that this is not in fact the case. A number of studies in recent years suggest that there exists a great deal of variability in the ocean interior at periods from one year to many decades. It is worth examining here the methodology that might be used to monitor such variability in more detail in the future.

It is well-known that variability related to the El Niño-Southern Oscillation (ENSO) phenomenon, while strongest at low latitudes, also appears in the subarctic N. Pacific. A number of studies have documented the effects of ENSO at higher latitudes. One of the most provocative of these studies (Jacobs et. al., 1994) recently suggested the possiblity that the 1982 ENSO event, the largest in recorded history, generated a Kelvin wave in the eastern Equatorial Pacific, which in turn propagated north along the North American coast and subsequently generated a Rossby wave at the western coast of North America that has continued to propagate across the N. Pacific, with the first arrivals of this wave impacting Japan and the Asian coast in mid-1993. This study suggested that, associated with this Rossby wave during the decade required for it to cross the N. Pacific, were sea level anomalies in excess of 25 cm and SST anomalies < 1°C, over horizontal length scales of approximately 1000 km. Much of the inference in this study is based on the results of numerical simulation, and it is fair to state that this study is presently being widely discussed in the oceanographic community and is somewhat controversial. Nonetheless, it is useful to consider the possibility of observing such an event from a monitoring system if it were to occur. The AVHRR, satellite altimetry, and hypothetical surface drifter array described above should all be able to discern such an event (indeed, the Geosat altimeter played an important role in the Jacobs et. al. study) at any given time. However, it is the duration of the event—over ten years were required for the leading edge of the Rossby wave to cross from North America to Japan—that poses the difficulty in monitoring phenomena such as these. To examine this event properly (assuming that it did indeed occur), it would have been desirable to have high quality AVHRR SST data, or SST maps from ships of opportunity, a satellite altimeter, and an array of about 60 surface drifters capable of measuring SST, operative over the entire decade from 1982–1993. Additionally, it would be desirable to process the data from these instruments in nearly real-time, so that the event could be studied as it evolved. The temporal and spatial sampling parameters of these measurement systems, and their measurement accuracy, are clearly capable of monitoring such events. Since in general these monitoring systems or their rough equivalents *were* in operation for at least part of the decade from 1982–1993, it is hoped that there will now be an incentive to return to the archives and search for the signature of this Rossby wave, with the goal in mind to verify in detail the hypothesis put forth by Jacobs et. al. (1994).

At even lower frequencies, there is now growing evidence that there can be substantial variability in the upper ocean. An outstanding example of this is given by Steven Hare of the University of Washington in his PhD dissertation (Hare, 1995; unpublished). This work shows dramatically that there has been a measurable long-term change in both the SST and the air pressure at sea level over the central North Pacific between 1947 and 1992. Over this time it is clear that springtime SST in the central N. Pacific has *fallen* by as much as 1°C, while the sea level pressure has fallen by as much as 6 millibars. Hare (and others; see Trenberth and Hurrell, 1994) suggests that this change in temperature is due to a long-term change in the position of the Aleutian Low center of action in the atmosphere. The essential explanation offered is that over time the wintertime position of the Aleutian Low has migrated eastward, towards the center of the North Pacific, causing a decrease in sea level pressure in the central N. Pacific. This has tended to intensify the southerly winds along the Pacific coast of North America and northerly winds in the central N. Pacific, leading to the presence of more *warm* water in the spring along the coast of N. America (and hence, an increase in temperature compared to the long-term mean value) and more *cold* water in the central N. Pacific (and hence a decrease in temperature compared to the long-term mean value there). The result is summarized in the anomaly maps shown in figures 3 and 4. What is perhaps most intriguing about this result is its temporal evolution, as shown in figure 4. There, it is seen that rather than undergoing a gradual change in position from 1947–1992, the Aleutian Low apparently changed its wintertime position and the springtime SST decreased rather abruptly in 1977, for reasons presently unknown. This is an extremely important observation, with implications to many problems in fisheries management and meteorlogical forecasting, and we are fortunate to have observed it so well. Yet how often do such events take place, and what are their space and time scales? Clearly it is exactly situations such as this one that we would like to to monitor, at the largest spatial scales (the Pacific Basin) and longest time scales (the duration of one scientist's career) that we can easily contemplate. The upper ocean monitoring system suggested above, including long-term satellite altimetry, long-term satellite based AVHRR derived SST, and a long-term array of approximately 60 surface drifters capable of measuring SST and meteorlogical parameters, could indeed discern the evolution of such an event if the observational system was put into place and continued in operation over a period of decades.

How well would the mean and standard deviation of SST need to be known in order to compare maps from year to year and discern significant changes in the "mean" SST fields over time for an event such as that shown in figure 3? If we suppose that we would like to know the mean SST at 500 km resolution, with a standard error of <.2°C (see the calculation in Section 2.1 above), then the sampling would have to continue for 4000 days, or approximately 11 years, in each 500 km square. In principle, this would allow the mapping of contours greater than .2°C in figure 3 with some statistical confidence, which would reveal the most important features of the change in SST over time. Of course, from equation (5) it is seen that the amount of data required decreases inversely with the square of the accuracy required; if it were deemed acceptable to know the mean SST field to only .4°C, then only 1000 days of data, about 2 1/2 years, would be required, and the largest anomalies in figure 3 would still be discernable by differencing the average SST maps over different time intervals. These would seem to be minimum sampling requirements, however, and higher spatial resolution (and the corresonding higher accuracy in mapping) is always preferable.

3.2. *The deep interior of the North Pacific*

Monitoring large-scale variations in the *subsurface* subpolar gyre would be a more difficult exercise than monitoring the sea surface, since satellite techniques such as altimetry and AVHRR could not be used. Furthermore, there is considerably less archival information available concerning the subsurface ocean that could be used for planning such an undertaking. However, even under these limitations it should be possible to produce periodic maps of the temperature and velocity fields using ALACE-type drifters (Davis et. al., 1992). These instruments are designed to drift at a preselected subsurface depth for a time of a few days to a few weeks and then to pop up to the sea surface, report their positon and temperature to an orbiting ARGOS satellite, to descend to back to their original depth, and to repeat this process approximately 40 times. If the time between ascents was set to 10 days, then an array of approximately 60 of these floats could be used to produce objective maps of the subsurface thermal and motion fields at approximately 500 km resolution. As with a sea surface experiment, such an experiment would have to continue for many years in order to determine the seasonally-dependent strength of the subsurface mesoscale eddy field with some degree of precision. The signal-to-noise ratio (ie, the ratio of mean energy to eddy variance) at depth in the N. Pacific is not well-known, but there is evidence from a few sites (see figure 5) that the ratio is somewhat depth dependent, with the most favorable signal-to-noise ratios at mid-depth around 1000 m. If the integral and square-integral time scales are similar throughout the water column (there is no good observational basis for this assertion), this implies that roughly the same amount of data will be required in space and time to discern the low-frequency behavior in the deepest portions of the ocean as is required at the sea surface, with perhaps somewhat less at mid-depths. The historical data base from the deep sea is sorely lacking, however, making experimental planning difficult.

One of the best examples of long-term subsurface variability in the N. Pacific interior has been given by Tabata (1989). From 27 years of monthly observations of temperature, salinity, dissolved oxygen, and nutrients from the sea surface to 5000 m at Station P in the Northeast Pacific (50°N, 145°W; approximately 2000 km from the North American continent), Tabata suggested that the two dominant time scales of variation were 2.5 years and the band of periods between 6–7 years. He attributed this variability to Rossby waves forced by the quasi-biennial oscillation of the atmosphere, but it is clear that it is also near the frequency band usually associated with ENSO effects. Additionally, at even lower frequencies, the work showed a long-term warming and freshening through the upper 1000 m of the water column. It was suggested in the work that the observed variability was characteristic of the region within a few hundred kilometers of Station P.

Tabata's results clearly demonstrate that a long-term committment must be made in order to properly monitor the deep-sea interior. Even after 27 years of observations the temporal statistics of the subsurface ocean were not stable, and it obvious that a record many years longer will be necessary in order to be able to examine the low-frequency behavior of the spectrum. If this is typical of the deep N. Pacific, then low-frequency spectral analysis of the N. Pacific will not be possible for a very long time, perhaps many decades, once a full-fledged monitoring program begins. While it seems useful to initiate such monitoring if resources and funds are available, the results from Station P suggest that constructing spatial maps from measurements distributed over a large number of independent points, such as might be measured from a large array of subsurface drifters, is likely to be a more productive exercise than attempting to estimate very low-frequency portions of the spectrum.

3.3. *Near-boundary regions of the N. Pacific*

The answers to many scientific questions will require monitoring of boundary regions of the N. Pacific over long periods of time. While the logistical problems of sampling near continental margins are simpler than the monitoring of the large-scale ocean interior, formidable problems exist in formulating a design for a coastal monitoring system. The spectrum of variablity in near-boundary regions may be as "red" as mid-ocean spectra and subject to considerable local variation, suggesting that the coastal ocean might need to be monitored over a period of time comparable to the long duration required for mid-ocean measurements.

3.3.a. *The Pacific coast of North America*

From 28 years of sea level data, Chelton and Davis (1982) showed that there exists a strong coherence of sea level at large scales along the coast of North America, from Acapulco to the Aleutians. Their study concluded that this variability was forced by variations in air pressure, with less significant contributions from ENSO-related effects. Of particular interest is their conjecture that the subtropical and subpolar gyres of the N. Pacific fluctuate out-of-phase, in that the temporal intensification of one gyre corresponds to a weakening of the other gyre (figure 6). This hypothesis is based on the inference that the coastal boundary currents (ie, the California Current and the Alaskan Stream) along the Pacific continental margin of N. America fluctuate in an out-of-phase manner due to an out-of-phase fluctuation in sea level along the coast, with the stagnation point taken to be somewhere slightly north of San Francisco. Somewhat similar and related conclusions have been drawn by Roden (1989), who used a smaller set of coastal sea level stations to examine variability along the Pacific coast of North America dating from 1827.

The inferences drawn from such large-scale, long-term variability of the coastal ocean by Chelton and Davis are convincing and show what might be accomplished with a few decades of sea level data from a large number of sea level stations extending over spatial scales of several thousand kilometers. The Chelton and Davis study used monthly mean data from 20 individual coastal stations. It is assumed that most of these stations will continue to be operational into the foreseeable future, and the agencies responsible for the operation of these stations and others like them are urged to continue these measurments. Beyond coastal sea level stations, *variations* in sea level can be measured from satellite altimeters at very high spatial resolution. There is, of course, potential difficulty in using an altimeter too near to land when the satellite trajectory is in the direction of land to ocean. This is due to the time necessary for the altimeter to come to equilibrium when first encountering the ocean (trajectories from ocean to land do not have this problem). In areas close to the continental margin, semidirurnal and diurnal tides are expected to be large, and this implies the necessity of having a tidal model that can be used to remove these signals from altimetric data where orbital tracks may repeat at intervals of a week or two.

Sea surface temperature is more difficult to measure at high spatial resolution in the coastal zone, and most studies to date have been forced to use composite data sets such as the COADS data. If it is truly desired to monitor the coastal ocean in real time, then these data sets will not be adequate. Instead, ships of opportunity or SST-measuring surface drifters will be required. The latter appear to be especially suitable for use in the coastal ocean. If it were desired to monitor SST over 4000 km of coastline (as in the Chelton and Davis study) at 200 km resolution alongshore and 50 km resolution offshore, and it were desired to monitor within 300 km of the coast (essentially the width of the coastal boundary currents), approximately 120 SST-equipped surface drifters would be required for the eastern N. Pacific. This would allow for nearly real-time maps of SST and surface flow to be constructed at intervals of a few days. This is an attractive possibility, and for an effort 3 times as large (360 drifters) the entire coastal region of the Pacific Rim, from Tokyo to San Francisco, could be mapped at intervals of a few days, in a fashion analogous to that suggested for the ocean interior in Sections 3.1 and 3.2 above. It would be necessary to maintain such an array for many years in order to adequately determine mean values, variances, and to discern long-term trends. To construct maps of temperature and velocity for the deeper waters of the nearshore ocean, approximately the same number of subsurface ALACE drifters (360) might be used along the Pacific Rim margin.

In some special regions, however, it is recognized the spatial coverage of a surface drifter array or a subsurface ALACE array will be inadequate to resolve features of interest. Such regions might include the Kuroshio system, plus coastal boundary currents such as the California Current of Alaskan Stream, both of which are considerably narrower than 50 km. In such regions it might be useful to maintain moored arrays of sensors that can be used to examine changes in these features and estimate their spectra. These moored arrays might resemble the Atlas moorings used so successfully in TOGA—TAO array in the Equatorial Pacific, or they might take some other form. While their maintenance over a long period of time might be considerably more costly than that of maintaining a large ensemble of drifters, such monitoring will be required in some crucial areas.

3.3.b. *Low-frequency temporal variations in coastal temperatures*

Royer (1989, 1993) has suggested in recent years that coastal ocean temperatures over a range of depths from 0 to 250 m at high latitudes in the subarctic N. Pacific show strong variability (in excess of 1.5°C) in the band of periods from 15–20 years, and he has hypothesized that this variation is somehow forced by the 18.6 year lunar nodal tide, although at the present time the observations span a time interval only slightly longer than one 18.6 year period (figure 7). It appears that local air temperatures at some stations, as well as precipitation, are highly correlated to these changes in ocean temperatures. Furthermore, it appears that the abundance of some fish stocks, such as Pacific halibut, are also correlated with these changes. Thus, it would seem to be useful to monitor the temperature of the coastal ocean from top to bottom, as well as related atmospheric variables, over a long period of time at a number of coastal stations, in order to test hypotheses concerning the causes of this variabilty. The temperature data used in Royer's study resulted from serial collection at a single station in the Gulf of Alaska at approximately monthly intervals over the period from 1971–1992.

It would appear to be somewhat difficult to examine this variability from a *spectral* point of view, since the frequency of variation is so low. It is desirable to have records 8–10 periods long in order to achieve some degree of statistical reliability, amounting to nearly 2 centuries of data in this case; however, being longer than the combined careers of 6 scientists, this seems to be a prohibitive requirement. It seems more prudent to proceed as Royer has done, continuing to collect the data over as long a duration as possible and to examine the data for long-term changes. In the long term, however, it seems possibly more cost effective to deploy a permanent, long-term telemetering mooring at the measurement site. In principle, atmospheric information could be collected at the same site. The *spatial* extent of this variability is unknown, and thus it would be useful to carry out such a measurement program and a number of sites in the subpolar gyre.

Beyond the verification that a very low-frequency, narrow band, temperature signal of significant amplitude exists at some scale in the subpolar gyre of the N. Pacific, it appears that other highly significant questions related to Royer's observations might be addressed from an experiment designed to monitor the subarctic coastal temperature. Most importantly, what is the source of this variability? The possible correlation with the lunar nodal tide cannot, in itself, be used to conclude that long-term variations in the lunar orbit can force changes in ocean temperature and, apparently, heat flux. Moreover, how much is known of the ocean's response to lunar nodal forcing? From an examination of spatial distribution of low-frequency variations of sea level at other long-period tidal frequencies, it has been concluded that at least some of the long-period tides are *not* in equilibrium (see Wunsch [1967] and Miller and Wunsch [1973]). As of yet, it is unknown whether or not the lunar nodal tide is in equilibrium, and it seems to be necessary to examine both sea level and ocean temperature signals in order to assess this question. At the present time, it appears that we are in general *not* able to address these questions, although the answers may have important societal impacts. Our best hope to make progress on these problems seems to be to initiate a monitoring program of the appropriate variables (sea level, depth dependent temperature, and atmospheric parameters) at a number of fixed sites in coastal subarctic N. Pacific over a long period of time. The spatial scale of this variability is unknown but possibly quite large, perhaps greater than 1000 km (n.b. the meridional scale of the lunar nodal forcing, according to Royer, is several thousand kilometers). Supposing this conjecture to be correct, then it would be useful to sample water column temperature, sea level, and atmospheric variables at approximately 1000 km scales. If, for example, it were desired to measure such variability over the continental margin of the Pacific Rim, this would require monitoring the ocean at approximately 10–15 sites from Japan to northern California; this sampling would need to be continued in perpetuity. For the next decade or so, it is suggested that the Gulf of Alaska be monitored as it has been for the past 20 years; if the data continue to show the strong 15–20 year variability discussed by Royer, then perhaps the apparent connection with the lunar nodal tide can be better understood.

3.3.c. *Variability in straits and passages*

An important oceanographic problem concerns the interaction of the N. Pacific with its marginal seas, such as the Japan, Okhotsk, and Bering Seas. One method of gauging such interactions is to monitor the exchange between the N. Pacific and the marginal seas through straits and passages such as exist in the Aleutians, the Kuriles, and between the islands of the Japanese archipelago. While such transports are in principle not difficult to measure by traditional means, in general the time dependence and evolution of such mass fluxes have been examined over time in only a few places. Reed (1990) was able to measure the exchange between the Bering Sea and N. Pacific through Amchitka Pass over a complete annual cycle at several depths; flows with speeds in excess of 40 cm/sec were observed, with fluctuations occurring on 30–60 day time scales. One of the drifters discussed by Stabeno and Reed (1991) sampled the flow through Chetvertiiy Strait between the Okhotsk Sea and the N. Pacific, with speeds of about 30 cm/sec through the strait. In Tsugaru Strait, between Hokkaido and Honshu, Dr. N. Shikama of the Japan Meteorlogical Agency has measured the transport over a period of two years using acoustic doppler (ADCP) methods. Curiously, the seasonal variations in transport in both Amchitka Pass and Tsugaru Strait are both rather small compared to the record mean values (no long-term measurements of transport are known to exist from the passes of the Okhotsk Sea).

These observations suggest that long-term monitoring of transport in such straits could be carried out using presently available current meter, sea level, and ADCP methods. In places where bottom telephone cables exist across straits, these could be used to monitor transport by electromagnetic methods. The ADCP and electromagnetic methods are especially attractive, since the possibility of *mechanical* failure is minimized. If temperature and conductivity measurements could be collected in conjunction with an ADCP program, then heat and salt fluxes through the most important straits and passages could also in principle be estimated. A potential pitfall, however, is the aliasing due to tides as discussed in Section 2.2 above. In many relatively shallow, narrow straits the tidal fluctuations (M_2 for example) can be sizable (50–100 cm/sec; see figure 8), sometimes even larger than the monthly mean value that is most useful for ocean monitoring. Thus, unless the relevant variables in the straits are sampled at a high enough rate that the tides are well-resolved (approximately every 6 hours), it is highly likely that the measurements will be degraded by undesirable aliasing.

4. DISCUSSION AND SUMMARY

The case for long-term monitoring of the ocean interior and the coastal zone is not difficult to make. This study has attempted to suggest simple sampling strategies that could be used to carry out such monitoring. It appears that the technology exists at the present time to adequately monitor temperature, sea level, velocity, and some important atmospheric parameters at the sea surface of the interior N. Pacific, although carrying out such an effort will be very expensive. It is suggested that much of the monitoring of the interior subarctic N. Pacific, either at the sea surface or at depth, could be carried out using an array of approximately 60 surface drifters or an equal number of ALACE drifters in the subsurface ocean. Such an array of drifters, either at the sea surface or in the deep sea, could be used to yield maps of the relevant variables at roughly 500 km resolution every 10–20 days. Floats and drifters such as these have been used extensively in WOCE and TOGA, and they have performed exceptionally well. If it is determined that it is necessary to measure the spectrum of variability at a few well-defined points, then at a limited number of locations in the subarctic gyre long-term moorings could be maintained, although at a higher cost than for the drifters. A similar drifter/float array, consisting of approximately 360 drifters in the surface ocean and ALACE floats in the subsurface ocean might be used to monitor velocity and temperature in the coastal ocean of the Pacific Rim, at scales of 200 km alongshore and 50 km offshore. Again, it is clear that it is desirable also to measure at a few fixed points (15–20 from Japan-California) in order to estimate the spectrum of variability over the long-term. Long-term measurements of exchange between the N. Pacific and its marginal seas, through straits and passages, are also of very high priority.

Monitoring the deep-sea will clearly be more difficult than monitoring the upper ocean, due to the important contribution made by orbiting satellites at the sea surface. At the present time, subsurface floats of various types, deep-sea current moorings, and acoustic tomography are the only technologies readily available for monitoring the subsurface ocean, and, while floats and current moorings are widely used in the oceanographic community, the utility of acoustic tomography has yet to be fully demonstrated. Each of these methodologies has its own inherent strengths and limitations for monitoring the deep-sea. In the future, however, new technologies, such as the *Slocum* drifter described by Henry Stommel (1989) may revolutionize our ability to sample the ocean beneath the sea surface. The *Slocum* concept of self-propelled vehicles, programmed to collect data while cycling between the sea surface and deep-sea along a preprogrammed path, and telemetering data to shore stations in near real-time over the course of a decade, is an extremely attractive one. Preliminary efforts are now underway to design and build such an instrument. When (and if) such devices become available in the future, monitoring the *entire* ocean, from top to bottom and from coast to coast, over the course of decades, should become a much more straightforward proposition.

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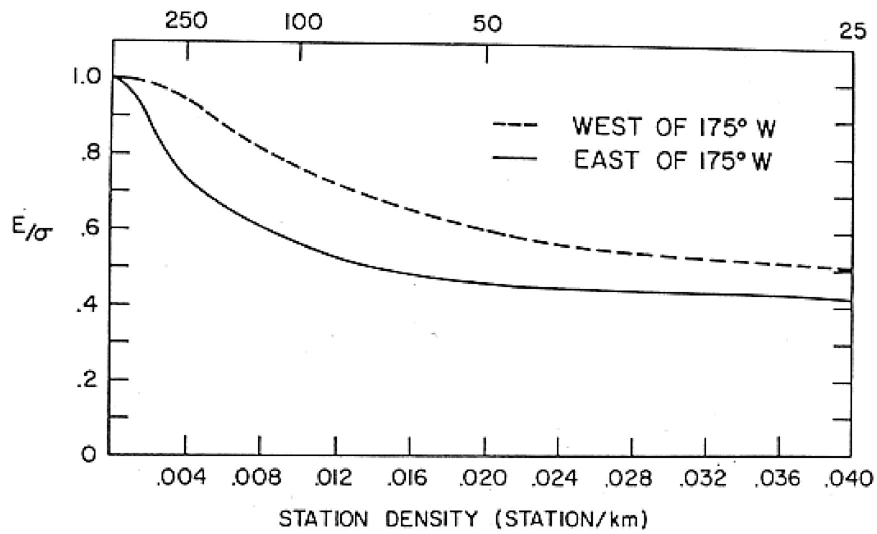

Fig. 1. The ratio of objective mapping error (E) to standard deviation of the large-scale upper ocean temperature field of the mid-latitude N. Pacific (σ) as a function of the spatial density of the sample points. From White and Bernstein (1979).

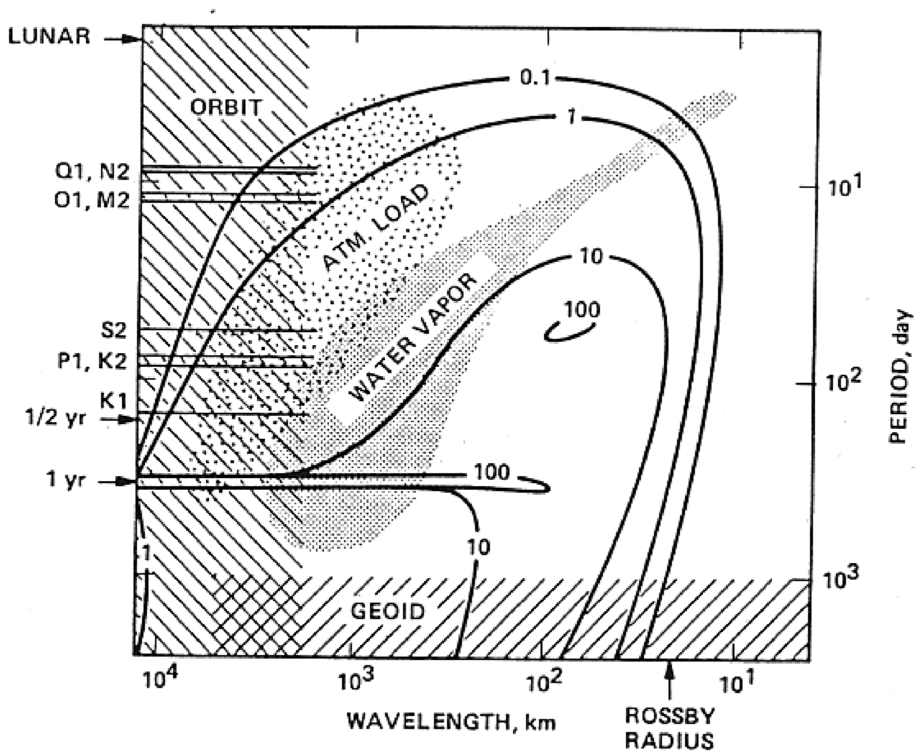

Fig. 2. A hypothetical frequency-wavenumber spectrum of sea level for the ocean at mid-latitudes, from Stewart et. al. (1986).

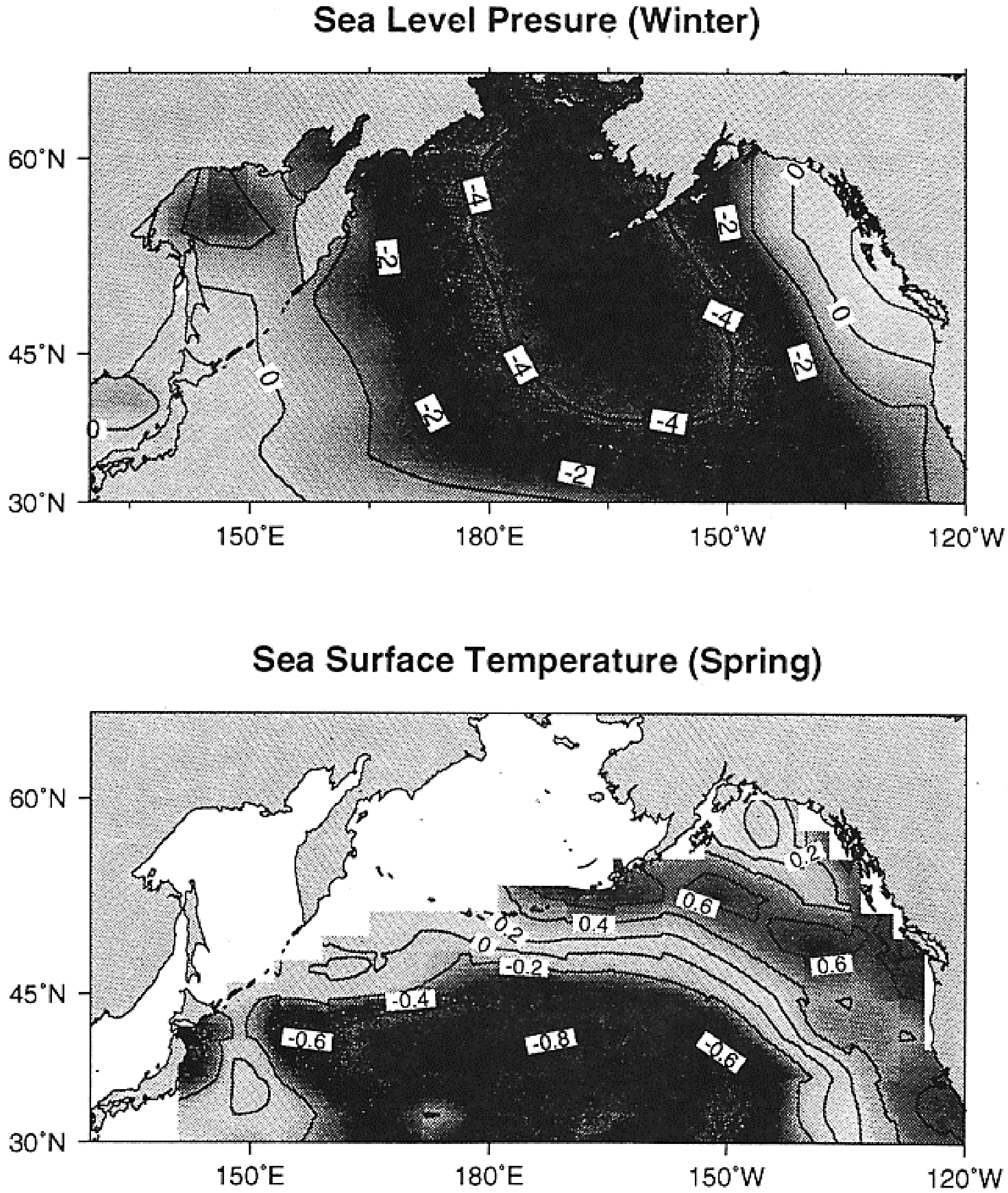

Fig. 3. The wintertime difference in COADS sea level pressure computed by subtracting the average sea level pressure during the period 1947–1976 from the average during the period 1977–1992.

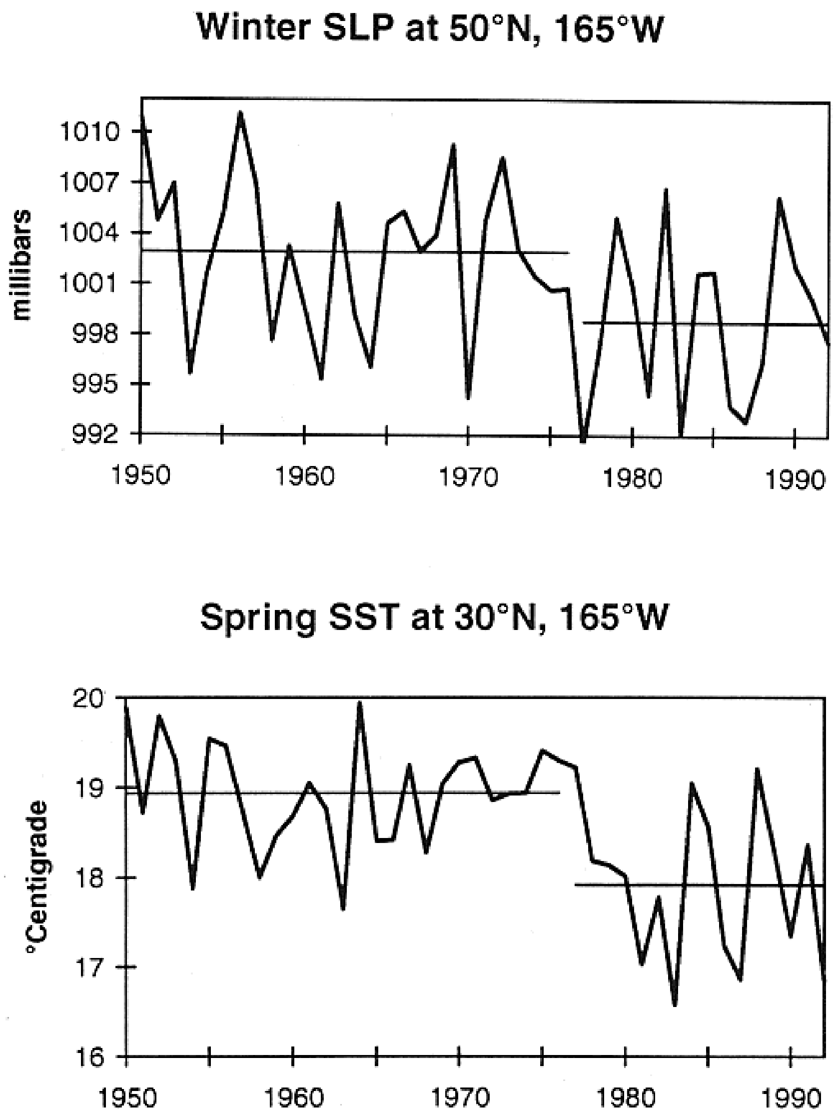

Fig. 4. Winter sea level pressure and spring SST at two sites in the subarctic N. Pacific, showing long-term changes. From Hare (1994).

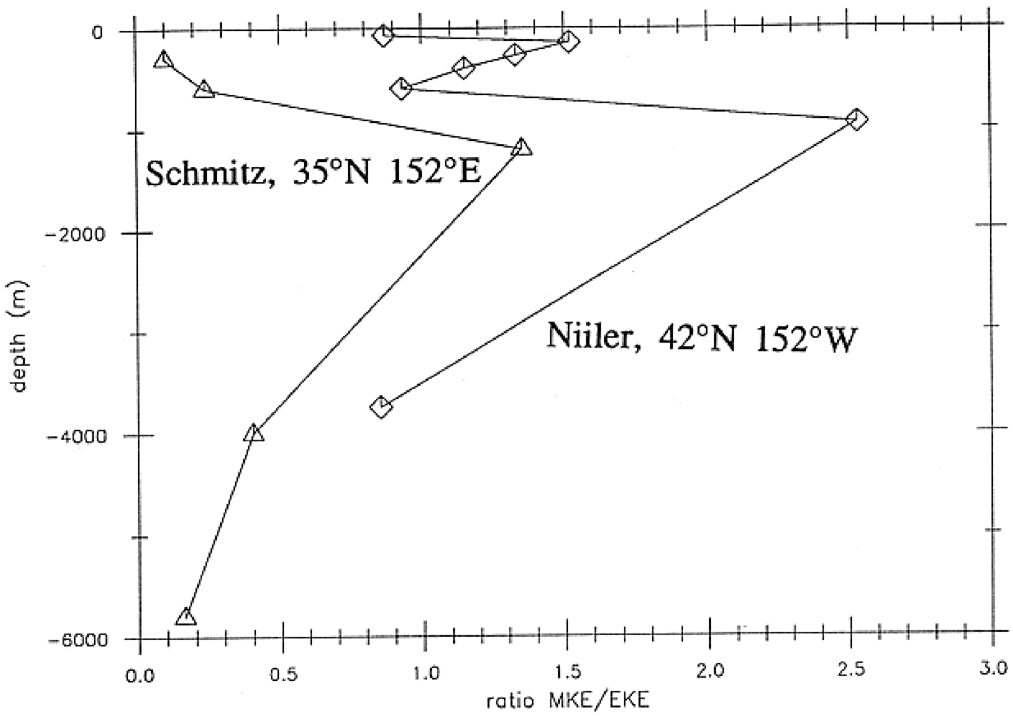

Fig. 5. The ratio of mean kinetic energy to eddy kinetic energy as a function of depth for 2 sites in the subarctic N. Pacific. These are estimates computed from the data of Schmitz et. al. (1987) and Hu and Niiler (1987).

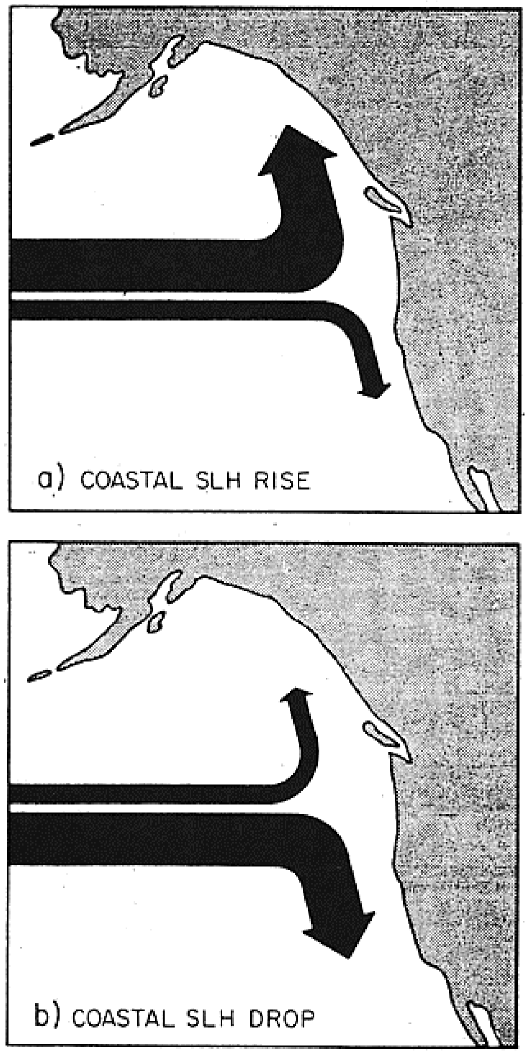

Fig. 6. A possible scenario for changes in coastal boundary currents related to changes in coastal sea level, from Chelton and Davis (1982).

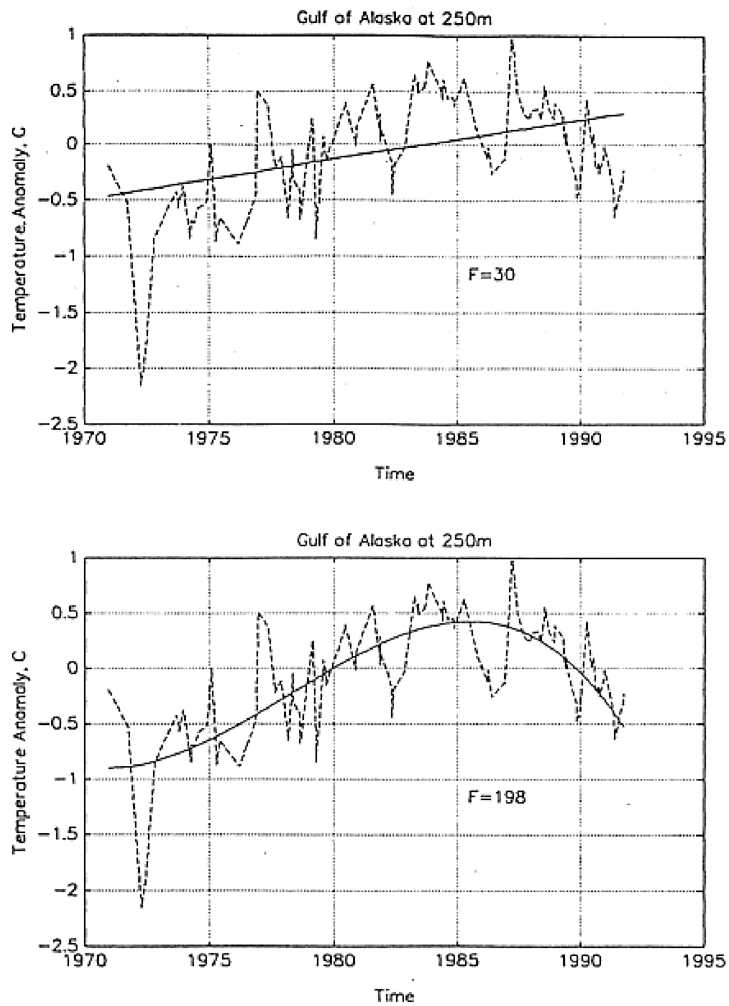

Fig. 7. Temperature at a depth of 250 m as a function of year at a site in the Gulf of Alaska, shown with a linear trend (top) and a nonlinear trend (bottom). From Royer (1993).

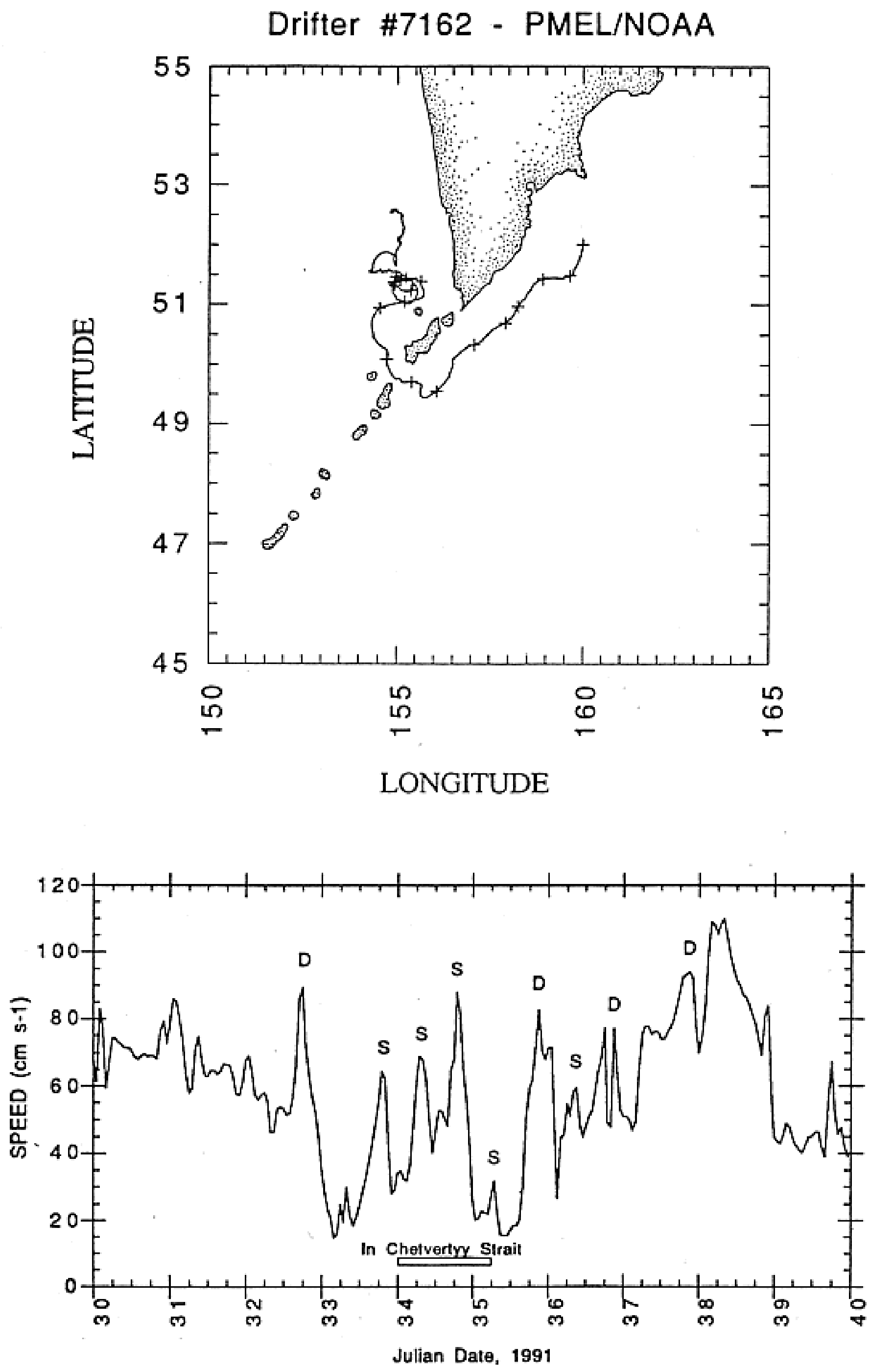

Fig. 8. Top panel: The track of a surface drifter entering the Okhotsk Sea from the North Pacific through Chetvertiiy Strait. Weekly positions are shown.

Bottom panel: The speed of drifter over a 10–day period as it passed near Chetvertiiy Strait, sampled a 2–hourly intervals. The peaks marked *D* and *S* denote features due to the diurnal and semidiurnal tides. Created using the data reported by Stabeno and Reed (1992).

Proceedings:
ECOSYSTEM RESPONSE

PICES-GLOBEC International Program on Climate Change and Carrying Capacity

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The North Pacific Marine Sciences Organization (PICES) and the Global Ocean Ecosystem Dynamics Program (GLOBEC) agreed in 1993 to jointly organize an international science program on "Climate Change and Carrying Capacity" (CCCC) in the temperate and subarctic regions of the North Pacific Ocean.

PICES is interested in studying Climate Change and Carrying Capacity due to the remarkable changes that have occurred in the North Pacific in recent decades, in both the open ocean and marginal seas. Concurrent changes in atmospheric pressure and ocean temperatures indicate that in 1976 and 1977 the North Pacific shifted from one climate state, or regime, to another that has persisted through the 1980s. Analyses of the North Pacific sea surface temperatures (SST) and atmospheric flow have identified a pattern of regime shifts in SST, the atmospheric Pacific/North American index, and the southern oscillation index, lasting several years to decades. Specifically, since 1976 there has been an intensification of the Aleutian low during the winter (November through March). The centre of the low has shifted further east and is now about 4 mb deeper, on average. There have also been associated changes in wind stress curl, the corresponding Sverdrup transport, a warming over Alaska, and a cooling in the central and western North Pacific. The strength of flows in the Alaska and California Currents also appear to fluctuate out-of-phase. Global modelling studies further suggest that if global warming occurs, its effects will be most strongly developed in high latitudes.

Although the important linkages are currently poorly understood, there is growing evidence that biological productivity in the North Pacific responds to these decadal-scale shifts in atmospheric and oceanic conditions, by flipping between periods of high and low productivity. In coastal areas, both the far eastern and California stocks of Pacific sardine peaked in abundance in the 1920s and 30s, declined significantly in the 1950s and 1960s, then began to increase synchronously in the mid-1970s. Large scale changes in pelagic fish production in the western Pacific, and in year class synchrony in recruitment of numerous important fish stocks in the eastern North Pacific, suggests coastal production is linked to variations in ocean climate. In the central North Pacific Ocean, there was a 50% increase in the average summer primary production, a two-fold increase in the summer biomass of zooplankton during the same period. The combined, national catches of salmon in the North Pacific also apparently declined steadily from historic highs in the late 1930s to a low in the mid-1970s. However, by the late 1970s there was a striking increase and the combined salmon catches subsequently had risen two- to three-fold, to nearly the historic high levels for this century. During any particular ocean climate regime the productivity of some species may be high, while the productivity of other species may be low. For example, major shifts in the dominant fish species has occurred among sardine, anchovy and mackerel in the Kuroshio-Oyashio Current region from the 1970s to 1980s. In addition to the decadal-scale regime shifts, longer-term global climate change may result in substantial changes in the biological carrying capacity of the North Pacific.

The CCCC Program will address how climate change affects ecosystem structure, and the productivity of key biological species at all trophic levels in the open ocean and coastal North Pacific ecosystems. The physical environmental changes that have occurred in this century, particularly during the late 1970s, may provide a natural experiment for studying such questions.

PICES-GLOBEC SCIENCE PLAN

The CCCC Program will take a broad ecosystem approach, with a strong emphasis on the coupling between atmospheric and oceanographic processes, and their impact on the production of the major living marine resources. Activities in the CCCC Program are anticipated on two spatial scales:

A. basin-scale studies to determine how plankton productivity and the carrying capacity for high trophic level, pelagic carnivores in the North Pacific changes in response to climate variations.

B. regional-scale, ecosystem studies comparing how variations in ocean climate change species dominance and productivity of key plankton and fish populations in the coastal margins of the Pacific Rim, from China to California.

The Science Strategy for the Program will include five main elements: (1) focus primarily on determining how the dynamics of the open subarctic and coastal margin ecosystems around the Pacific Rim respond to climate change, (2) employ mechanistic process studies to improve understanding and develop early recognition and prediction capabilities for regime changes, (3) develop and employ models to guide research activities, integrate results, and improve capabilities for forecasting ecosystem responses to climate change, (4) develop broader insights and understanding through regional scale comparative studies, and (5) support and coordinate CCCC Program activities with GLOBEC.INTernational and other existing and planned international (e.g. NPAFC, WOCE, JGOFS, GOOS, NOPACCS) and national (e.g. CalCOFI, BIOCOSMOS, GLOBEC-SPACC, HUBEC, FOCI, LaPerouse) organizations and research programs in the PICES region.

The Key Research Activities proposed include:

1. Retrospective analyses of existing atmospheric, physical, plankton, and fisheries data, to identify recent (and historical) changes in the subarctic Pacific.

2. Develop numerical modelling system for ecosystem dynamics research and monitoring.

3. Develop observation and monitoring systems to collect new observational data on surface winds, upper ocean temperature and circulation, nutrient flux, phytoplankton productivity (from ocean colour sensors and in situ studies), zooplankton, mesopelagic micronektonic animals and higher trophic level carnivores.

4. Biological Process Studies will be conducted in the open subarctic, and in the selected suite of regional, coastal ecosystems around the Pacific Rim. Key process studies will focus on primary production, zooplankton production, distribution and life history, food web trophodynamics and modelling (feeding, growth, reproduction and mortality rates), life history models, and the migratory behaviour of key zooplankton, fish, marine mammals, and sea birds.

PICES-GLOBEC International Program on Climate Change and Carrying Capacity - An update on progress

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The PICES-GLOBEC Workshop held in Nemuro on 15-17 October 1994 led to agreement on a Science Plan which was subsequently approved by the Governing Council of PICES and published in the report of the PICES Third Annual Meeting. A Scientific Steering Committee for the CCCC Program was established and charged with development of an implementation plan. Membership of that Committee was established and its Executive Committee met in Honolulu on 24-26 May 1995.

The Draft Implementation Plan that emerged identified several Central Scientific Issues:

Physical Forcing: What are the characteristics of climate variability, can interdecadal patterns be identified, how and when do they arise?

Lower trophic level response: How do primary and secondary producers respond in productivity, and in species and size composition, to climate variability in different ecosystems of the subarctic Pacific.

Higher trophic level response: How do life history patterns, distributions, vital rates, and population dynamics of higher trophic level species respond directly and indirectly to climate variability?

Ecosystem interactions: How are subarctic Pacific ecosystems structured? Do higher trophic levels respond to climate variability solely as a consequences of bottom up forcing? Are there significant inter-trophic level and top-down effects on lower trophic level production and on energy transfer efficiencies?

Research activities applied to these issues at both basin and regional scales would include (1) retrospective analyses, (2) development of models, (3) process studies, (4) development of observation systems, and (5) data management.

The following regional (1-10) and basin scale (11-12) components of the CCCC Program were identified; a tentative list of national and interna-tional programs, present or planned, in each component was prepared.

1. California Current System, south
2. California Current System, Oregon to Vancouver Island
3. Southeast and Central Alaska
4. Eastern Bering Sea
5. Western Bering Sea/Kamchatka
6. Okhotsk Sea
7. Oyashio/Kuroshio

8. Japan Sea/East Sea
9. Bohai and Yellow Sea
10. East China Sea
11. Western subarctic gyre
12. Eastern subarctic gyre

Of particular importance for the regional scale studies is a comparison of ecosystem properties and responses to climate variability. For such comparisons, a common set of program outputs is required; the following items were proposed:

Physical forcing: location of major fronts and current boundaries; atmospheric pressure gradients; air-sea heat exchange; major physical features; mixed layer temperature and depth; velocity of major currents; eddies, vertical and horizontal mixing and fine structure.

Lower trophic levels: annual and seasonal productivity; temporal and spatial patterns of plankton dynamics and nutrient fields; identifica-tion of major taxonomic groups; population parameters for key species or taxonomic groups.

Higher trophic levels and ecosystem interactions: abundance trends and distributions of life stages of key species and their predators and prey; population parameters (growth, mortality, reproduction); food web structure (including diets and trophodynamic linkages of key species); pro-duction and productivity structure.

It is proposed that the Scientific Steering Committee now be called the Implementation Group. Because of the importance of the regional scale studies, representatives of National GLOBEC Programs constitute the majority of the Group's Executive Committee which also includes a representative of the North Pacific Anadromous Fish Commission, with which PICES is co-operating in this program. Liaison has been established with SCOR and IOC, sponsors of GLOBEC International, and with the International Council for the Exploration of the Sea which is sponsoring a related program, on Cod and Climate Change. Cooperation with other relevant international programs and organizations is also contemplated.

The importance of monitoring to the CCCC Program is clear. The goal of the program is to understand, and eventually to predict, the effects of climate variation on ecosystems of the subarctic Pacific. This requires that changes in climate and in ecosystem response be measured systematically in an appropriate monitoring system, the design of which draws on the findings of retrospective, process, and model studies. Thus the PICES Working Group on Monitoring (WG 9) will have a key role in reviewing existing and planned monitoring systems and in taking the requirements of the CCCC Program into account as they develop plans for
enhancement of monitoring in the region.

Monitoring of Low Trophic Level Variability and Response to Long-Term Forcing in the Subarctic Pacific Ocean

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ABSTRACT

The subarctic Pacific Ocean consists of the Alaskan gyre and the western Subarctic gyre with a couple of seasonally ice-covered marginal seas such as the Bering Sea and the Sea of Okhotsk. The western Subarctic Gyre is the unique water mass, which can be modified by trapping of a warm core ring, in the subarctic Pacific Ocean. The Alaskan Gyre and the western Subarctic Gyre show a strong seasonal thermocline. Winter mixing does not occur below 100 m in the former gyre while it occurs deeper than 200 m in the latter gyre. This difference in the physical structure of water mass in winter may cause a different low trophic level variability at the near surface ecosystem between two water masses, and the low trophic level may respond differently to long-term forcing. To monitor those variabilities and responses, ship observation has been the main effort based on either continuous or discrete determinations. Some determinations, for examples, microzooplankton are still dependent mostly on the discrete sampling. Technological developments with acoustics, optics, and satellite should be achieved to overcome the present difficulties encountered in the subarctic Pacific Ocean.

1. SUBARCTIC PACIFIC OCEAN

The subarctic Pacific Ocean is isolated by the subarctic front from the other parts of Pacific Ocean (Roden 1972) and has a variety of water masses (Favorite et al. 1976). The Bering Sea and the Sea of Okhotsk are isolated from the subarctic Pacific Ocean by the Aleutian Islands and the Kurile Islands, respectively. They in part are seasonally ice-covered and will be discussed by somewhere else. The entire subarctic gyre has a positive Ekman vertical velocity (Gargett 1991). This indicates that a wind-forced upwelling replenishes nutrients in the euphotic zone. The eastern and the western parts of the subarctic Pacific Ocean are consist of the Alaskan Gyre, which encircles the Gulf of Alaska (Musgrave 1992), and the western Subarctic Gyre, which encircles the western subarctic Pacific Ocean (Kawai 1972), respectively. In the present study the term of the western Subarctic Gyre is applied to a broad area including the western subarctic domain defined by Dodimead et al. (1963). The western Subarctic Gyre often traps a warm core ring, which modifies the regional water structure (Kawasaki 1992). The physical structure of the Alaskan Gyre is characterized by a permanent halocline at 80-120 m (Tabata 1975). The annual exchange through the halocline is limited (Acara 1964, Tabata 1965) and relatively insignificant (Denman and Gargett 1988). On the other hand, a permanent halocline similar to the one in Alaskan Gyre dose not exist at a shallow upper layer in the western Subarctic Gyre (Favorite et al. 1976). Nutrients are abruptly transported to the upper layer in winter.

Even though the subarctic Pacific Ocean concerns this study, the difference in water structures between two gyres, which are outside the effects of any land mass or continental shelf, within the ocean, requires a specific monitoring system for each gyre to evaluate the response to the long-term forcing in this region. Most data to present the Alaskan Gyre in the present study are obtained at the oceanic station P (50°N, 145°W) and its vicinity. However, data to present the western Subarctic Gyre are obtained from station A11 (40°30'N, 146°00'E), which located the western edge of the western subarctic domain, but still at the north of the subarctic boundary (Kawasaki 1992).

2. LOW TROPHIC LEVEL VARIABILITY

The driving force for the surface water currents in the Alaskan Gyre and the western Subarctic Gyre is the drag of the wind on the surface of the water (Fig. 1) although there is the argument on the drag of the wind (Amorocho and DeVries 1980). Wind directions are opposite between winter and summer in the region of the subarctic Pacific. This variable force may affect the current speeds in the gyres since the geostrophic circulation is weak in the subarctic gyre. However the general current system remains the same regardless of wind directions. A strong seasonal thermocline is developed during the heating period in both gyres. It occurs in May in the Alaskan Gyre (Tabata 1975) and in April in the western Subarctic Gyre (Kasai et al. 1994). Surface winter minimum temperature was higher than 5.5°C in the former gyre (Fig. 2) and higher than 3°C in the latter gyre (Fig. 3). Vertical mixing occurs in winter in both gyres. However, vertical mixing dose not occur below 100 m in the former gyre (Tabata 1975) while it goes deeper than 200 m in the latter gyre (Kasai et al. 1994). Winter nitrate concentrations in the surface mixed layer are 13 uM in the Alaskan Gyre (Anderson et al. 1969) and 20 uM in the western Subarctic Gyre (Kasai et al. 1994), respectively. These values are reduced to a minimum of 6 uM in late summer in the former gyre. However they are reduced to less than 0.5 uM in the latter gyre (Fig. 4). Winter chlorophyll *a* concentrations are 0.2 mg m^{-3} in the Alaskan Gyre (Parsons and Anderson 1970) and 0.05 mg m^{-3} in the western Subarctic Gyre (Kasai et al. 1994), respectively. The former value increases to only 0.73 mg m^{-3} in summer (Welschmeyer et al. 1993) while the latter value increases to higher than 5 mg m^{-3} in the spring bloom (Fig. 5). The seasonal change of chlorophyll *a* concentration shows a single peak in summer in the Alaskan Gyre and one large in spring and one small peak in fall in the western Subarctic Gyre. According to the seasonal studies by Parsons and LeBrasseur (1968), Anderson et al. (1969), and Parsons and Anderson (1970), the timing of the spring bloom is earliest along coast and latest in a central area associated with the center of Alaskan Gyre. The similar trend is also observed in the western Subarctic Gyre (Taguchi et al. 1994, Kasai et al. 1994, Yoshimori et al. 1995). In the Alaskan Gyre, picoplankton (<2 um) is usually dominant beneath the seasonal thermocline in August (Miller et al. 1988). These small cells prefer the regenerated ammonium to nitrate (Harrison and Wood 1988). During the spring bloom, larger cells than 10 um become the most abundant; more than 90% of total chlorophyll *a* in the western Subarctic Gyre (Fig. 6). These large cells assimilate nitrate effectively (Taguchi et al. 1992).

Based on the ecosystem model, Frost (1993) described little seasonal change of phytoplankton standing stock in the Alaskan Gyre with a much large short-term variability. However, a study on numerical model of spring bloom reveals a great seasonality in the western Subarctic Gyre (Yoshimori et al. 1995). Seasonal change in phytoplankton production rate is substantial in the Alaskan Gyre (Frost 1993). Maximum daily primary production can exceed 1 gC m^{-2} and minimum values are below 0.2 gC m^{-2} with a modelled annual production of 162 gC m^{-2}. Although no seasonal data for primary production are completed at the present time, the similar substantial seasonal change in phytoplankton production is expected in the western Subarctic Gyre.

Zooplankton was collected with a vertical tow of 350 um mesh net from 150 m to the surface in the Alaskan Gyre (Parsons and Lalli 1988, Brodeur and Ware 1992). Macro- and microzooplankton were collected with a vertical tow of 333 um mesh net and 183-35 um mesh net from the bottom of surface mixed layer in the western Subarctic Gyre (Saito et al. 1994a). In the Alaskan Gyre there is a marked increase in the biomass of macrozooplankton as measured by wet weight (converted to dry weight by using a factor of 0.1) from winter values of about 1 mg m^{-3} to spring maximum of about 8 mg m^{-3}. On the other hand, average dry weights of macrozooplankton vary a range from about 5 to 400 mg m^{-3} in the western Subarctic Gyre. A ratio of the maximum to the minimum is 80, which is ten times higher than one in the Alaskan Gyre. Although seasonal change of microzooplankton biomass is not available in the Alaskan Gyre, LeBrasseur and Kennedy (1972) provide the seasonal change of microzooplankton community in numbers as an index of microzooplankton abundance. They show that maximum number of 30,000 individuals m^{-3} in winter and minimum number of 100 individuals m^{-3} in summer. However, these values seems low compared to <10 x 10^6 ciliates m^{-3} (Landry et al. 1993). Strom et al. (1993) estimated ciliate carbon biomass using a conversion factor of 0.19 pgC um^{-3} (Putt and Stoecker 1989). The ciliate biomass ranged from 1 to 7 mgC m^{-3} during summer. Simulated annual cycle of herbivorous carbon biomass ranged from <10 to 40 gC m^{-3} (Frost 1993). While herbivores is not the major consumers of phytoplankton (Dagg 1993), the specific mortality rate caused by microzooplankton is balanced by the specific growth rate of phytoplankton during spring and early summer (Landry et al. 1993). In the western Subarctic Gyre, microzooplankton carbon biomass changes from maximum of 100 mg m^{-3} in spring to minimum of 1 mg m^{-3} in winter (Saito et al. 1994b). Similar to the Alaskan Gyre, the ecological importance of microzooplankton compared to macrozooplankton is significant in summer in the western Subarctic gyre (Fig. 7). However, maximum zooplankton biomass, which occurred at or immediately after the spring bloom of phytoplankton, mostly consisted of macrozooplankton in the western Subarctic Gyre.

Neocalanus plumchrus, N. cristatus, and *Eucalanus bungii* may comprise 80 to 95% of the plankton biomass collected by 0.33 mm net in the Alaskan Gyre (Miller et al. 1984). They are particularly abundant in the spring and early summer at Station P. *N. plumchrus, N. flemingeri, N. cristatus,* and *E. bungii* may comprise more than 95% immediately after the spring bloom in the western Subarctic Gyre (Saito et al. 1994b).

Chaetognaths are important among invertebrate carnivores in the Alaskan Gyre (Terazaki and Miller 1986) and the western Subarctic Gyre (Terazaki et al. 1995). *Eukronia hamata,* which inhabits the epipelagic (0-200 m) and upper mesopelagic layer (200-500 m) is the most abundant species in the Alaskan Gyre. Seasonal maximum and minimum abundances are about 1.5 and 0.5 individuals per m^3, respectively. *Sagitta elegans,* which inhabits the epipelagic layer, is the most abundant species in the western Subarctic Gyre.

Carbon flux determined by sediment trap experiments (Takahashi 1986) is estimated at 200 m by fitting an exponential decay curve to data at great depths as studied by Pace et al. (1987). The maximum flux occurs in July and the annual carbon flux is estimated as 86 gC m^{-2} $year^{-1}$ in the Alaskan Gyre (Taguchi 1992). Carbon flux determined by sediment trap experiments at 500 m (Hanaoka, unpublished data) shows one small peak (30 mgC m^{-2} d^{-1}) in spring and one large peak (100 mgC m^{-2} d^{-1}) in fall in the western Subarctic Gyre (Fig. 8). Annual carbon flux at 1,145 m is estimated to be 12 gC m^{-2} $year^{-1}$ in the western Subarctic Gyre if 40 mgC m^{-2} day^{-1} was assumed for the period from May 12 to July 10 in which data was lacking. Annual carbon flux at 50 m estimated by the method of Pace et al. (1987) is 120 gC m^{-2} $year^{-1}$. When the concept developed by Eppley and Peterson (1979) is adopted to both gyres, the annual carbon flux can be converted to the annual primary production. The f-ratio is defined as a ratio of new production to new plus regenerated production. Annual new production estimated by Wheeler (1993) based on 180 day sufficient light for primary production is 49 gC m^{-2}, which corresponds to 30 % of modelled annual primary production (Frost 1993). This estimate is similar to the annual minimum estimated by Frost (1993). In this study 40%, as suggested by Frost (1993) is adopted to the Alaskan Gyre while 50% is applied to the western Subarctic Gyre since their primary production is more likely controlled by nitrate than ammonium (Kasai et al. 1994). When those values are applied to both gyres, the annual primary production is estimated to be 215 gC m^{-2} $year^{-1}$ in the Alaskan Gyre and 240 gC m^{-2}

year^{-1} in the western Subarctic Gyre, respectively. The former estimate is 1.3 times higher than the estimate from ^{14}C experiments (170 gC m^{-2} y^{-1}) (Welschmeyer et al. 1993). The independent estimate of annual primary production at Station P by Wong (cited in Welchmeyer et al. 1993) is similar (140 gC m^{-2} y^{-1}). When we consider error involved in those calculation, the difference between those two estimates can be insignificant but needs to be studied.

3. RESPONSES OF LOW TROPHIC LEVEL TO LONG-TERM FORCING

The western Subarctic Gyre locates between 40 and 55°N while the Alaskan Gyre locates between 50 and 60°N. Photosynthetically Available Radiation (PAR) should be similar between both areas (Sverdrup et al. 1965). However, more than 60% cloud coverage occurs with 60% of chance in the western Subarctic Gyre and higher than 70% of chance in the Alaskan Gyre in July (Fig. 9). This difference may result in different level of PAR on the ocean surface. Photosynthesis of phytoplankton may respond to the different level of PAR, which is governed by a long-term climate change, to achieve higher primary production in the western Subarctic Gyre.

Occurrence of low pressure system is limited to the western Subarctic Gyre in summer (Fig. 10). This may disturb temporarily a seasonal thermocline. However, the low pressure occurs often over the two areas in winter. Vertical mixing of surface water would be strongly induced by wind in winter. It is very critical for the supply of nutrients from the deep water in the western Subarctic Gyre. However the amount of nutrient supply is dependent on the different water structures between two gyres. The permanent halocline in the Alaskan Gyre could reduce the amount of nutrient supply from deep water. The upper ocean stratification above the permanent halocline may be related to a relatively higher precipitation in the Alaskan Gyre than that in the western Subarctic Gyre (Terada and Hanzawa 1984) and freshwater input as summer runoff from coastal areas along the eastern and northern boundaries of the Alaskan Gyre (Royer 1982).

Phytoplankton would respond to assimilate nutrients when a favorite condition comes in spring (Sverdrup 1953). The response of phytoplankton is different between two gyres. Large diatoms respond immediately to high nitrate concentrations in the western Subarctic Gyre while picoplankton assimilate nutrients in the Alaskan Gyre. Picoplankton may assimilate favorably ammonium regenerated by microzooplankton in the Alaskan Gyre as shown in the oceanic surface water by Harrison and Wood (1988). Large diatoms may not be efficiently grazed by macrozooplankton but they sink to deep water (Riebesell 1991). However, in the western Subarctic Gyre, macrozooplankton increase immediately after the increase of large diatoms (Fig. 11).

Bakun (1973) indicates that the strong interannual variation in the intensity of the winter winds is positively correlated with the average summer zooplankton biomass during 1956-1962 and 1980-1989 (Brodeur and Ware 1992). An increase in the intensity of the zonal winds associated with the North Pacific High could speed up the subarctic current and advect zooplankton into the southern Gulf of Alaska (Brodeur and Ware 1992). In the western Subarctic Gyre, the zonal index indicates a different trend in the spring of 1991 (Fig. 12) when the spatial distribution of abundant zooplankton biomass (>1.25 mgC m^{-3}) is limited to the coastal water and sporadic locations (Fig. 11). The high anomaly observed in March 1991 may be related to the reduced growth of young copepodites in spring.

4. RECOMMENDATION FOR LONG-TERM MONITORING

Effective methods are partially listed for the long-term monitoring in the subarctic Pacific Ocean (Table 1). Technology for most measurements is available at the present time. However, some application should be considered to monitor the long-term variability of lower trophic levels in the subarctic Pacific Ocean. Such application should be made (1) to determine the biological processes in the regenerated nutrient predominant ecosystem in the Alaskan Gyre, since preferential utilization of ammonium by phytoplankton can account for the low standing stock of phytoplankton and the underutilization of nitrate in the surface layer in summer, and (2) to determine the biological processes in the large particle predominant ecosystem in the western Subarctic gyre.

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Table 1. Effective methods for the long-term monitoring in the subarctic Pacific Ocean. ST and SP indicate the satellite and ship observation including buoy system, respectively.

Measurements	Information Obtained	Method	References
Wind	vertical mixing	ST, SP	Yan et al. (1991)
PAR	primary production	ST, SP	Kiefer et al. (1989)
PAR	heating	ST, SP	Archer et al. (1993)
Temperature	mixed layer depth	ST, SP	Yan et al. (1991)
Salinity	vertical mixing	ST, SP	
Nutrient	new production	ST, SP	Sathyendranath et al. (1991)
Current	spatial distribution	ST, SP	Fiedler (1984)
Shear	vertical distribution	SP	Gargett & Osborn (1981); Caldwell et al. (1985)
Natural fluorescence	chlorophyll *a* primary production	ST, SP	Kiefer et al. (1989)
Absorption	chlorophyll *a*	ST, SP	Moore et al. (1992)
Upwelling irradiance	primary production	ST	Gordon et al. (1983)
Upwelling irradiance	taxonomic group	ST	Bricaud et al. (1988)
Laser	size distribution	ST, SP	Tsuda et al. (1989)
Acoustics	biomass size distribution	SP	Napp et al. (1991)
Optics	biomass	SP	Herman (1992)
Sediment Trap	vertical flux	SP	Eppley & Peterson (1979)

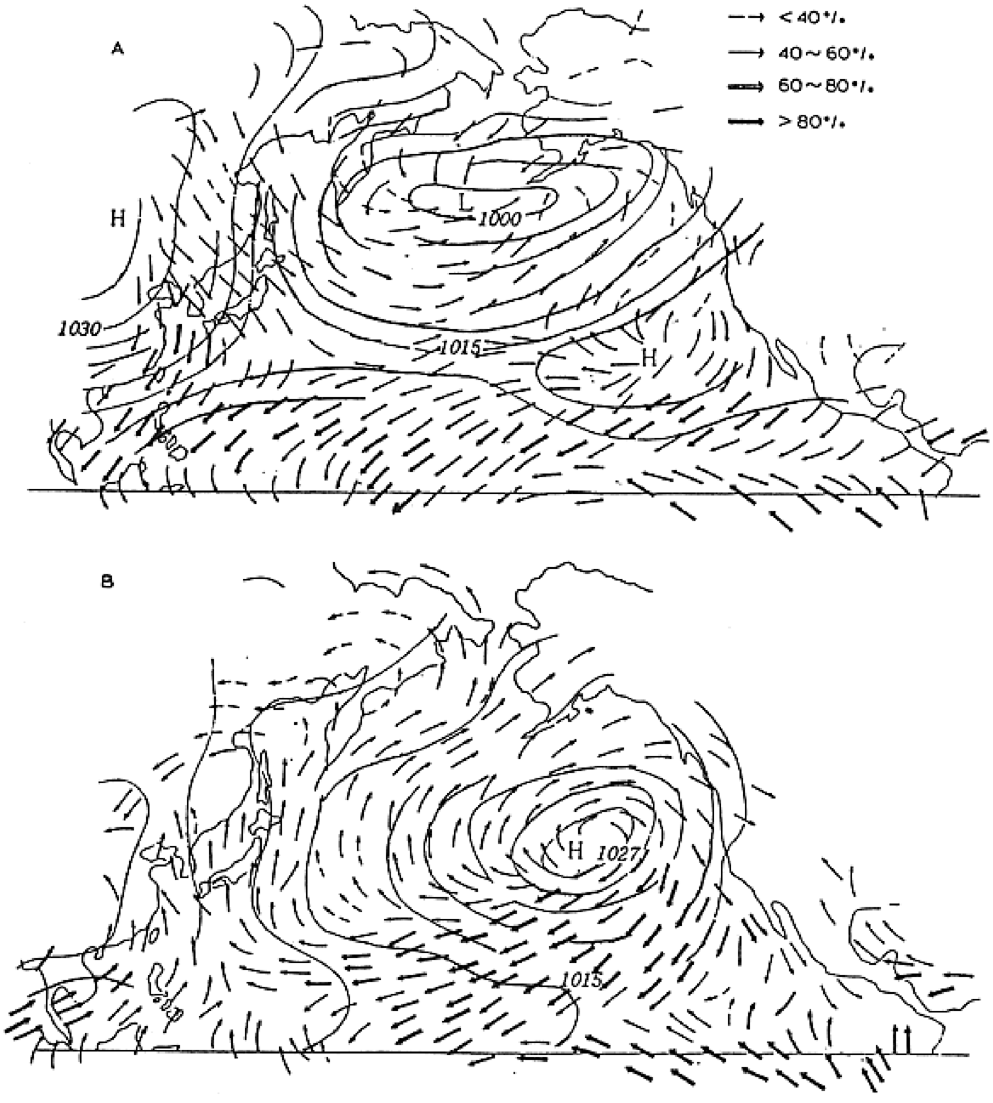

Fig. 1. Frequency of wind directions in winter (A) and summer (B). (after Terada and Hanzawa 1984).

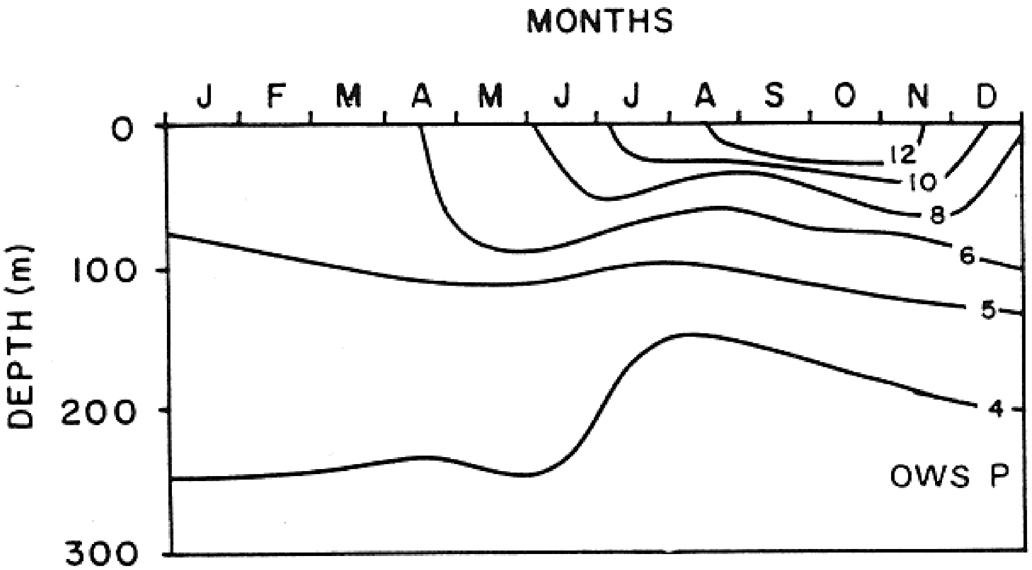

Fig. 2. Seasonal change in temperature at St. P (50°N, 145°W) in the Alaskan Gyre. (after Parsons and Lalli 1988).

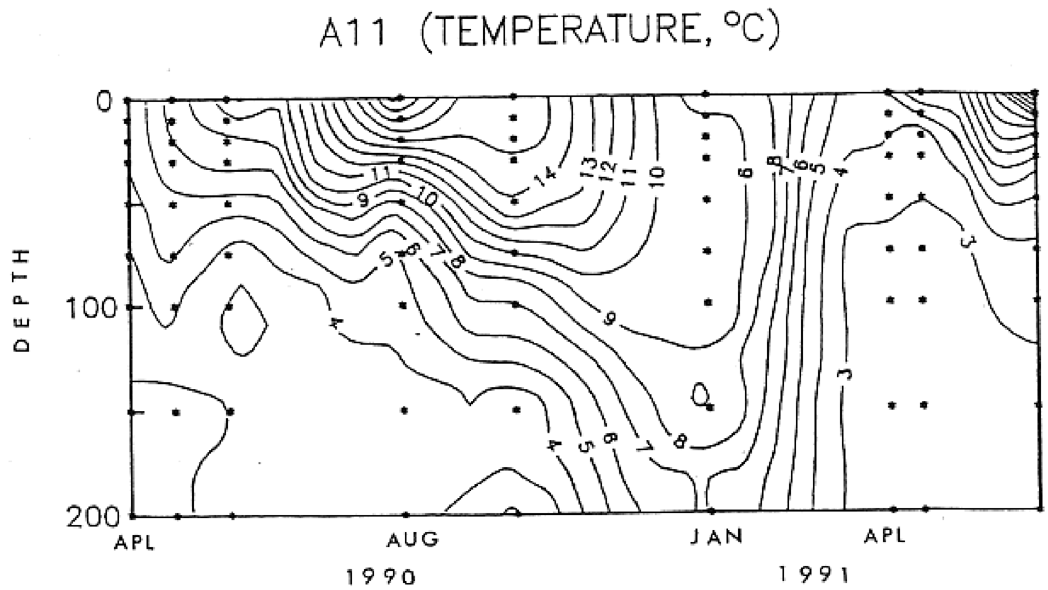

Fig. 3. Seasonal change in temperature at St. A11 (40°30'N, 146°E) in the western Subarctic gyre. (Kasai, unpublished data).

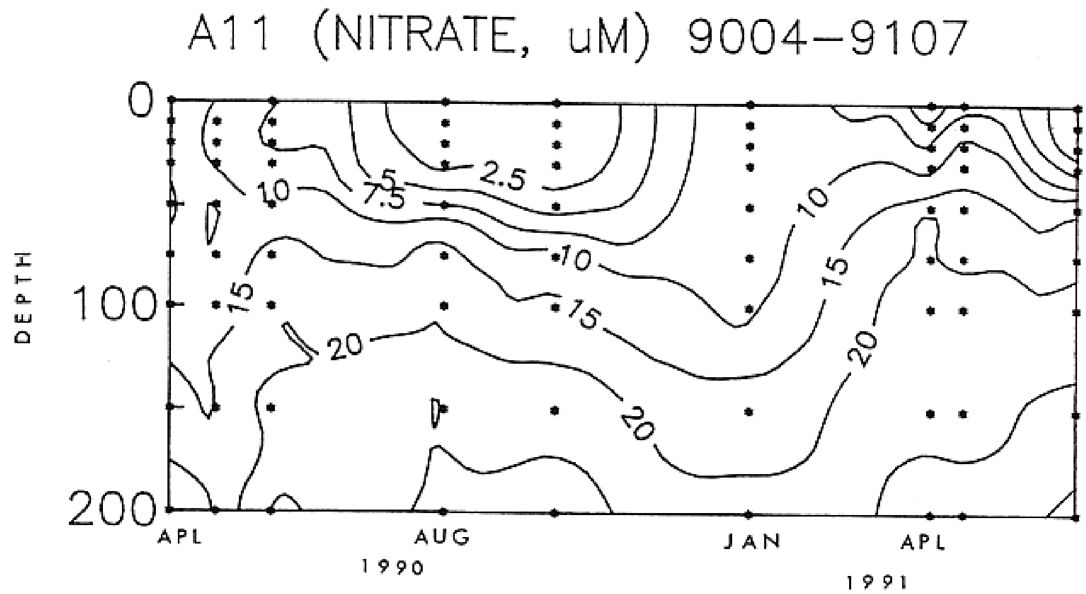

Fig. 4. Seasonal change in nitrate concentration (uM) at St. A11 in the western Subarctic gyre. (Kasai, unpublished data).

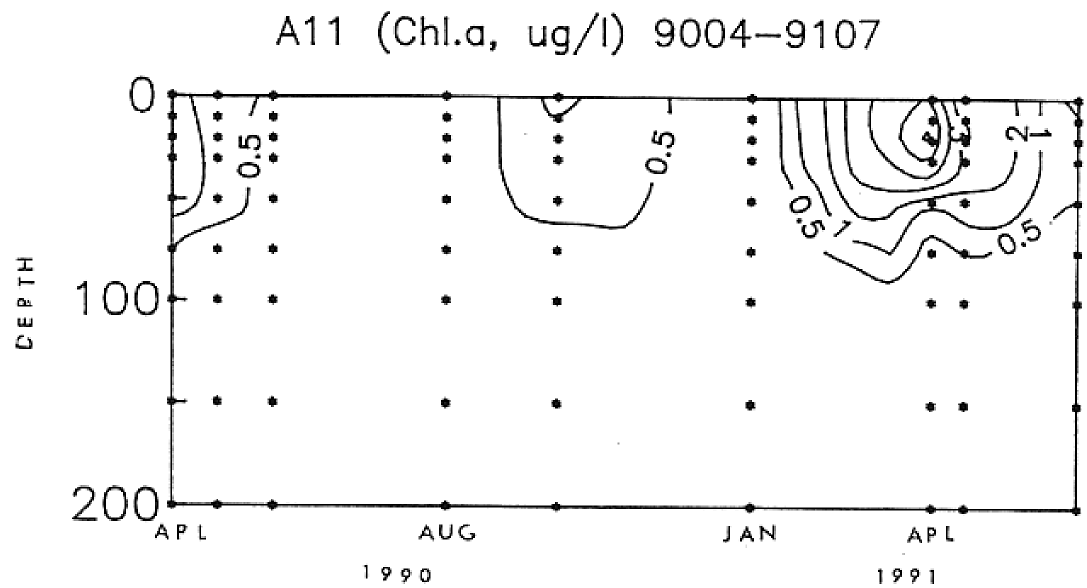

Fig. 5. Seasonal change in chlorophyll *a* concentration (ug l^{-1}) at St. A11 in the western Subarctic gyre. (Kasai, unpublished data).

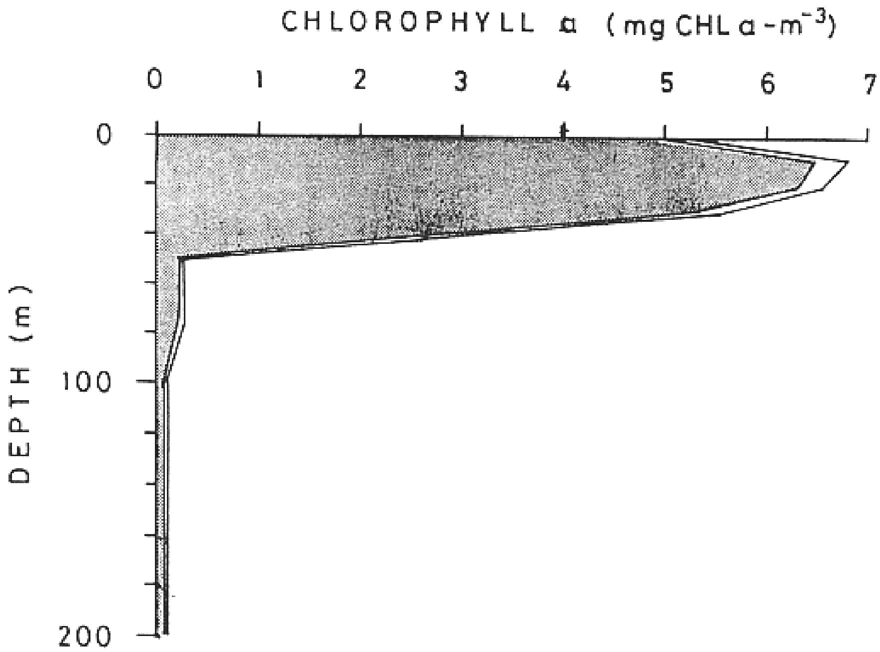

Fig. 6. Vertical distribution of chlorophyll *a* (mg m^{-3}) at St. A11 on June 5, 1990 in the western Subarctic gyre. Shade area indicates larger cells than 10 um.

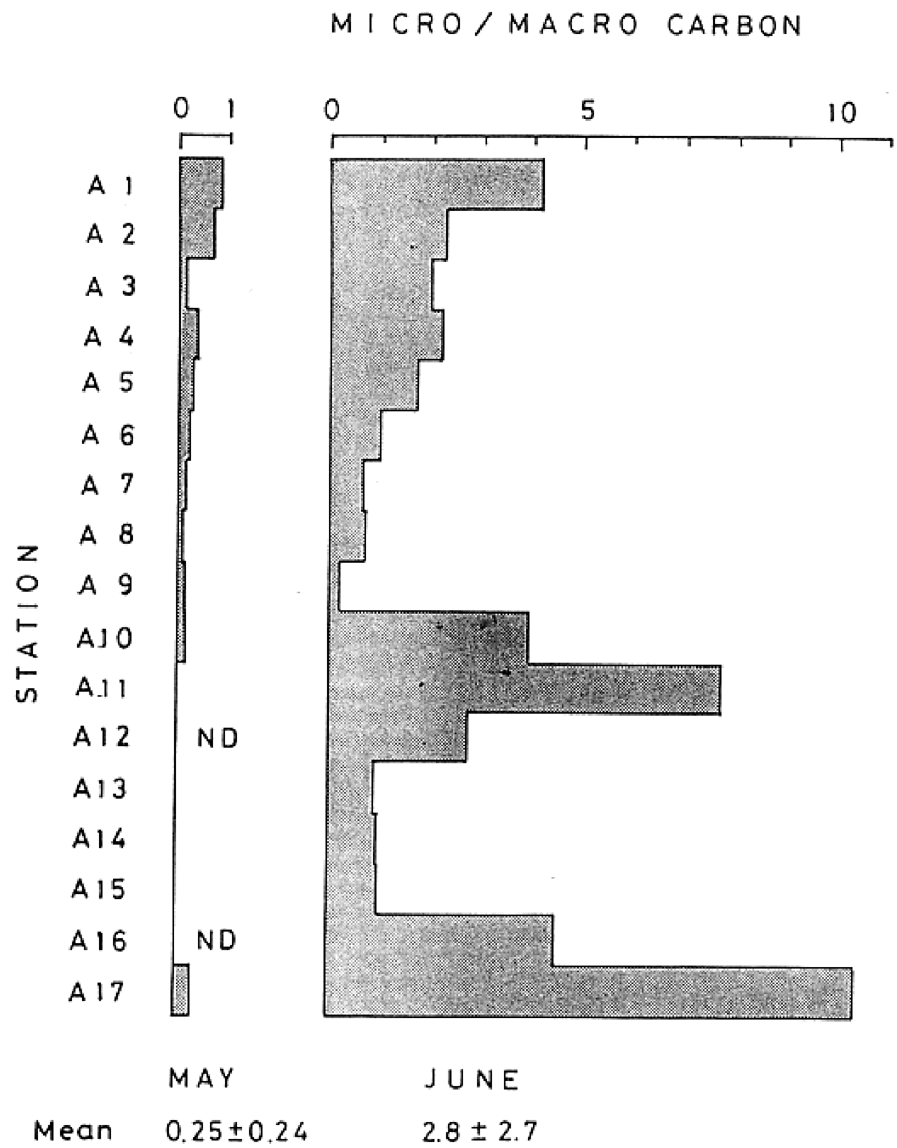

Fig. 7. A ratio of microzooplankton to macrozooplankton carbon in May, which corresponds to the spring bloom, and June along the oceanographic observation line off southeast of Hokkaido. A1 locates at 42°50'E, A17 locates at 39°N, 146°45'E.

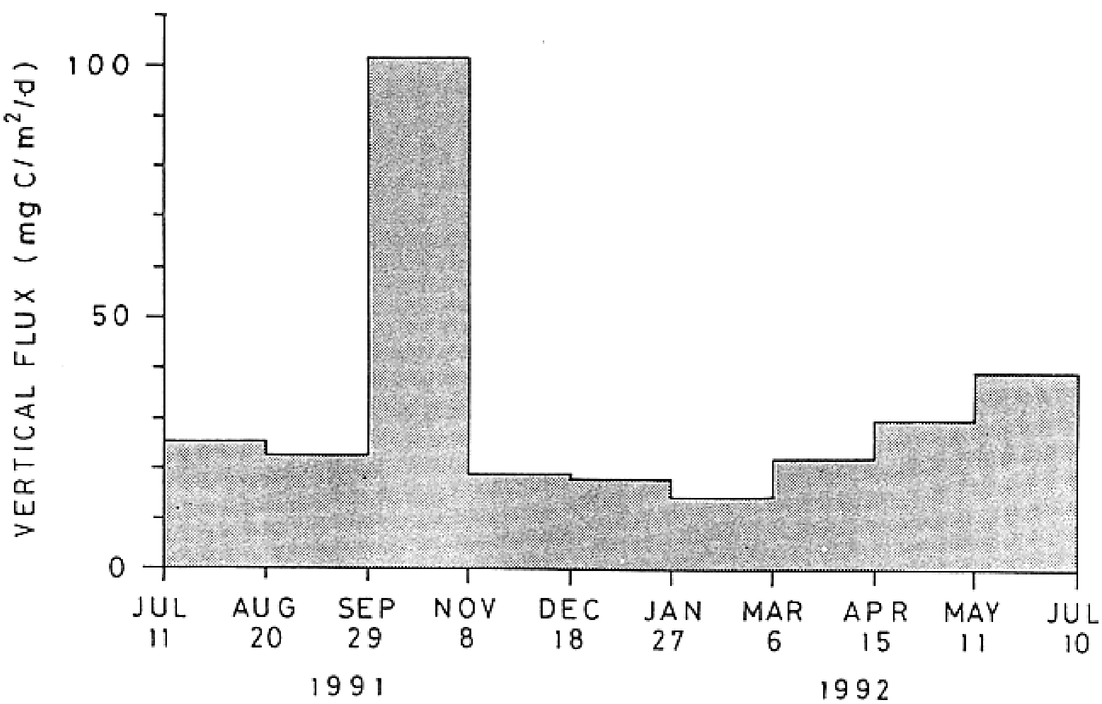

Fig. 8. **Seasonal change in the vertical flux of carbon off the southeast Hokkaido. (Hanaoka, unpublished data).**

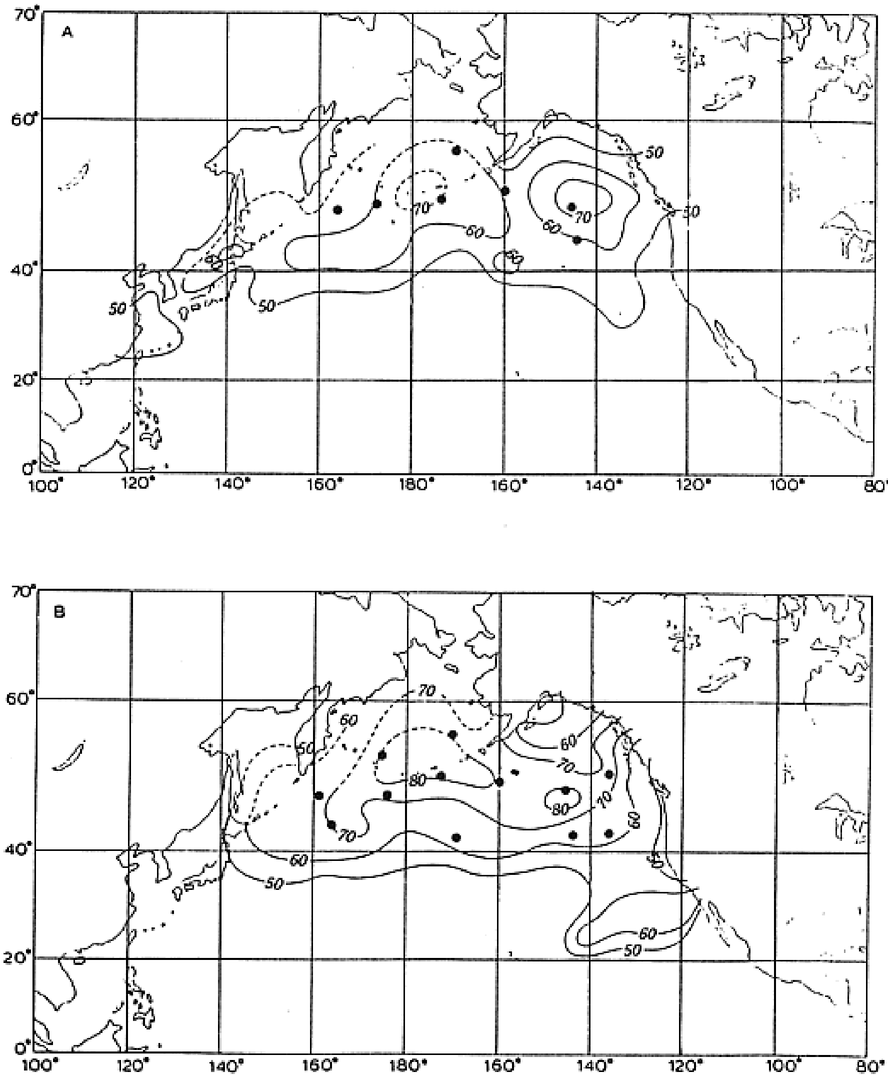

Fig. 9. Percentage frequency of total cloud amount higher than 60% in January (A) and July (B). Dotes indicate an average coverage higher than 85%. (after Terada and Hanzawa 1984).

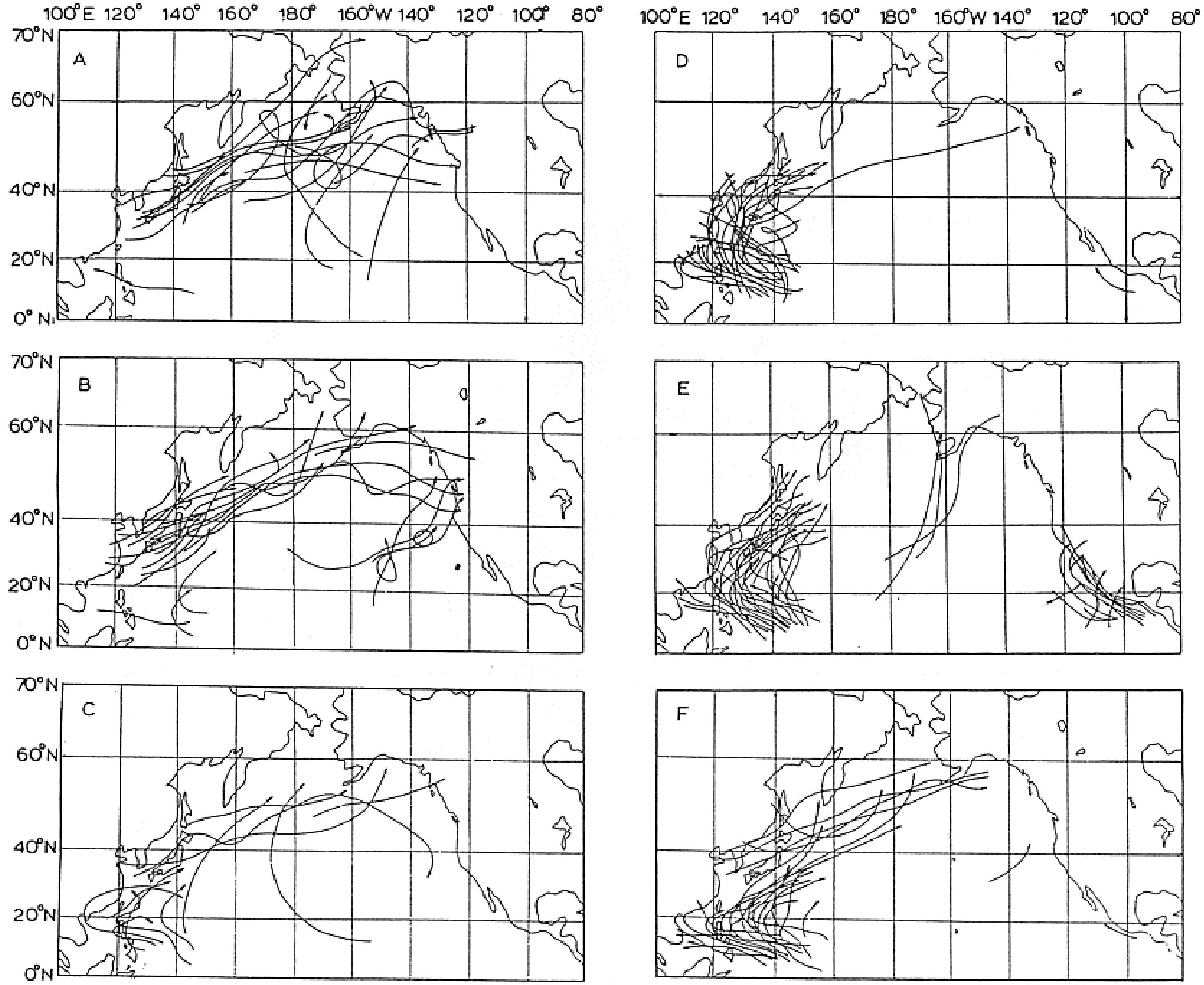

Fig. 10. Tracks of tropical depressions in January (A), March (B), May (C), July (D), September (E), and November (F) in the North Pacific. (after Terada and Hanzawa 1984.)

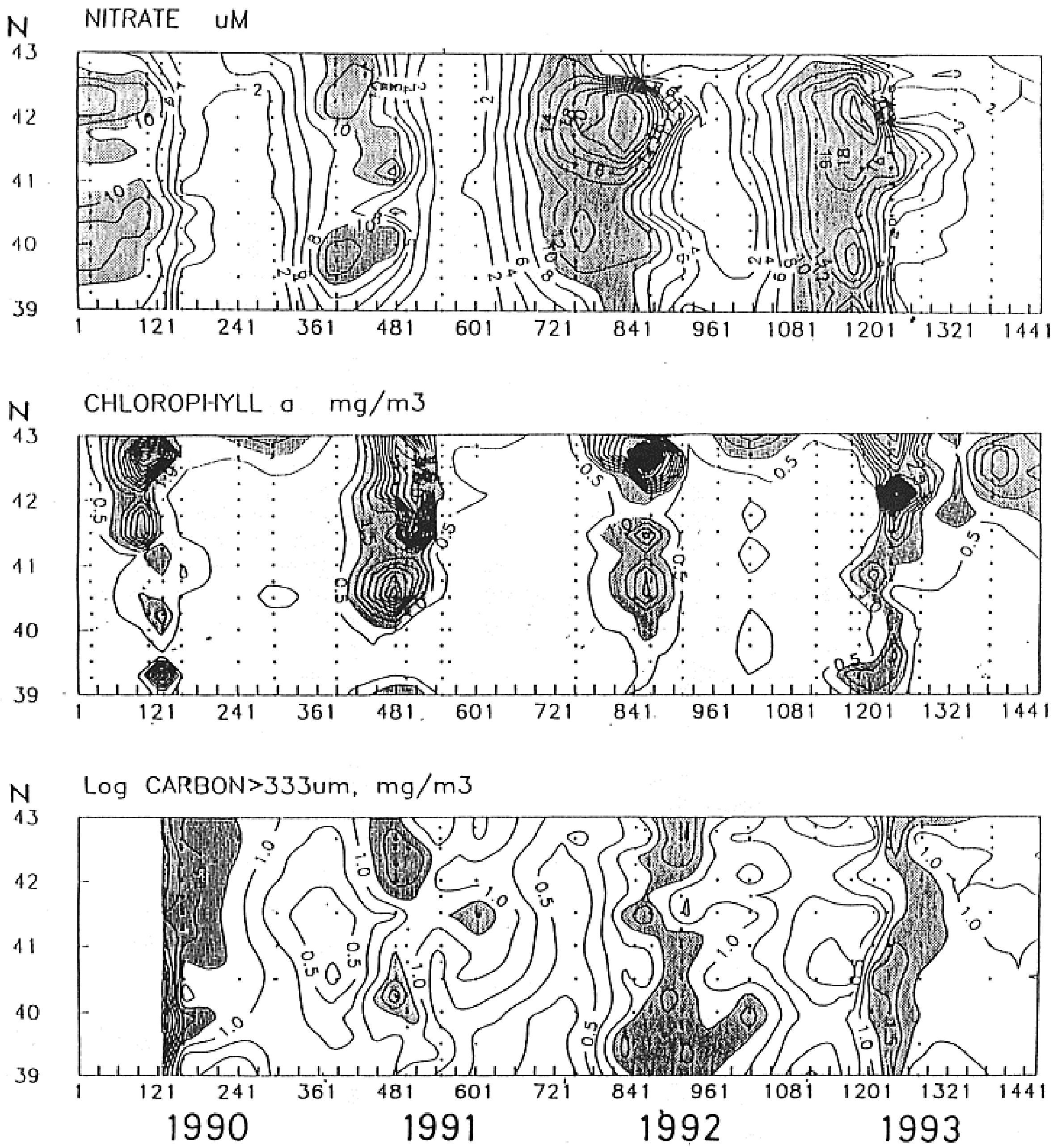

Fig. 11. Time-series observations of surface nitrate concentration (upper panel), surface chlorophyll *a* concentration (middle panel), and macrozooplankton carbon in the surface mixed layer (bottom panel) along the oceanographic observation line off the southeast Hokkaido during the period from 1990 to 1993. Shade areas indicate higher nitrate concentration than 10 uM (upper panel), higher chlorophyll *a* concentration than 1 mg m^{-3} (middle panel), and higher macrozooplankton carbon than 1.25 mgC m^{-3} (bottom panel), respectively.

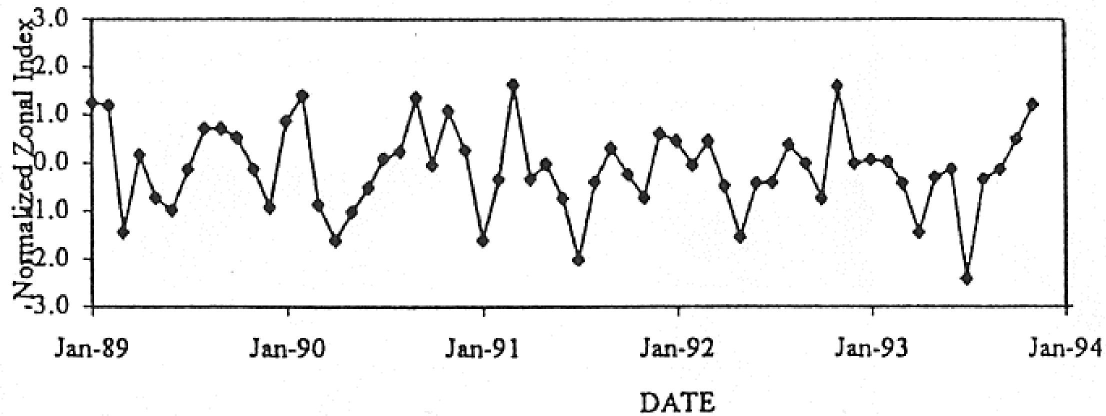

Fig. 12. Normalized zonal index calculated during the period from 1989 to 1993 in the western Subarctic gyre. (Saito, unpublished data).

RECOMMENDATIONS

1. General Findings

i) Some aspects of the climate-scale physics and biology of the subarctic North Pacific are understood, and long-term monitoring programs can be scientifically designed based on this knowledge and on existing technology. The climate module of the Global Ocean Observing System (GOOS) is a target for any recommendations concerning monitoring for the purposes of describing and understanding climate. GOOS, in addition to providing the ocean component of the Global Climate Observing System (GCOS), is a target for any recommendations concerning monitoring for the purposes of understanding climatic effects (namely, variability and change) on the living marine resources in the ocean, especially at the lower trophic levels.

ii) Other aspects of the subarctic North Pacific require more research, or new technology, before monitoring programs can be undertaken. GCOS and GOOS are not targets for these recommendations, because GCOS and GOOS are not research programs or funding agencies, however it would be useful to keep GCOS and GOOS informed as to research and technology needs and programs, so that close contact can be initiated and maintained to facilitate a future transition to monitoring programs.

2. Specific Findings and Recommendations

I) The working papers prepared for this Workshop are, in general, review papers of high quality and broad interest. They should be published so as to provide the material in archive and accessible form.

It is recommended that PICES publish the review papers as a PICES Scientific Report, after the authors have had an opportunity (two months) to revise their papers based on the discussions at the Workshop. The report will be restricted to the seven invited authors, plus a summary statement at the beginning by the co-editors, Prof. Sugimori and Dr. Briscoe. Specific information on formats and text requirements will be provided to the authors by the co-editors.

PICES should treat the recommendations and discussions of this Workshop as only the first step regarding monitoring that is possible now and the needs for new science studies and technology development. Therefore, it is recommended that PICES form an interdisciplinary, Monitoring Working Group with the following terms of reference:

- The WG will be responsible for planning the monitoring activities in the PICES area, including proposing scientific and technical priorities and schedules, and including physical, biological, and chemical measurements.

- The WG should cooperate with the GCOS Ocean Observing Panel for Climate, the GCOS Living Marine Resources science planning group, the Scientific Steering Committee of the PICES-GLOBEC *Climate Change and Carrying Capacity Program*, and other such bodies as may be needed.

- The WG will work with the PICES Technical Committee on Data Exchange to ensure timely and open exchange of monitoring data between participants and to external data users, as a mechanism to control the quality and relevance of the data.

- The WG will report regularly as requested by the PICES Science Board.

  The WG will have a lifetime of two years, which may be extended by the Science Board if needed. Its membership will be two people from each member country, with credentials in the scientific and technical areas concerned with the monitoring activities. The Chairman will be named by the PICES Science Board.

ii) New technology is essential to effective monitoring of the PICES area, but its development requires attention for a long period, often with not visible results, and at great cost. Nevertheless, it is essential.

It is recommended that PICES encourage its member states to support the development of the new technology required for monitoring and identified in the report of this Workshop. Particular attention should be paid to the development of autonomous biological instrumentation, without which monitoring of ecosystem response to climate forcing will be particularly difficult.