Assimilation of Satellite Altimetry into a Western North Pacific Operational Model

Masafumi Kamachi[1], Tsurane Kuragano[1], Noriya Yoshioka[2], Jiang Zhu[3], and Francesco Uboldi[4]

[1]*Meteorological Research Institute, 1–1 Nagamine, Tsukuba 305–0052 Japan*

[2]*Japan Meteorological Agency, 1–3–4 Otemachi, Tokyo 100–8122 Japan*

[3]*Institute of Atmospheric Physics, Chinese Academy of Science, Beijing 100029,*

[4]*Magritte SNC, Milano, Italy*

(Received September 1, 2000)

ABSTRACT

An ocean data assimilation system, COMPASS–K (the Comprehensive Ocean Modeling, Prediction, Analysis and Synthesis System in the Kuroshio–region), has been developed at the Meteorological Research Institute (MRI). The purposes of the development are understanding ocean variability in the Kuroshio region as a local response to a global climate change with assimilated four–dimensional data sets, development of an operational system in the Japan Meteorological Agency, and for the GODAE (Global Ocean Data Assimilation Experiment) project.

The model is an eddy permitting version of an MRI–OGCM. Space–time decorrelation scales of ocean variability are estimated with TOPEX / POSEIDON (T / P) altimeter data. Subsurface temperature and salinity fields are projected from the T / P altimeter data with a statistical correlation method and are assimilated into the model with a time–retrospective nudging scheme.

Seasonal variation in the western North Pacific is investigated. Realistic space–time distribution of the physical quantities, the path of Kuroshio and its separation from Honshu are captured well. The Kuroshio volume transport is well reproduced in a reanalysis experiment of 1993. Preliminary predictability experiments are done in February and March, 1994. Predictability diagram shows the time scale of the predictability for temperature field is about 17 days in the Kuroshio south of Japan. This time scale is smaller than that in the North Atlantic.

Key words: Assimilation, Kuroshio, Predictability

1. Introduction

Linear dynamics is dominant as a response to atmospheric forcing in the equatrial region. In the mid– to high–latitudes, ocean represents nonlinear phenomena such as strong currents and meso–scale eddies. Heat and water fluxes are also important. The resultant scales of the phenomena are rather small. We developed, for the mid– to high–latitudes, an ocean data assimilation system COMPASS–K: Comprehensive Ocean Modeling, Prediction, Analysis and Synthesis System in the Kuroshio region (Kamachi et al., 1998).

Data assimilation is a series of technical procedure combining observation data and a model with subtracting useful information from observation. The aims, generally, are obtaining an optimum initial condition for prediction, model improvements, understanding of oceanic phenomena using a coherent representation (e.g., 4D–movie), which is a result of the assimilation, of an evolving ocean system, and constructing an optimum observing network as a result of the sensitivity of the each data set.

The purposes of the development of the COMPASS–K are understanding ocean variability in the Kuroshio region as a local response to a global climate change with assimilated four–dimensional data sets, development of an operational system in the Japan Meteorological Agency (JMA) for nowcasting and forecasting of ocean state, and for a prototype system of the GODAE (Global Ocean Data Assimilation Experiment) project under GOOS (Global Ocean Observing System). This system is a comprehensive system and consists of ocean general circulation model (OGCM), analysis components (i.e., pre–analysis, quality control, assimilation scheme, optimization scheme, analysis of output), and prediction. Figure 1 shows the relationship of those components schematically.

In this system, TOPEX / POSEIDON altimetry and ship data are assimilated into the model in the North Pacific. TOPEX / POSEIDON altimeter data are important for the data assimilation, because the data have conditions: (1) data acquisition in near real time, (2) continuous (or periodic) acquisition and (3) wide (or global) coverage. Remote sensing data serve the surface information but not subsurface one, then we need a statistical method or a advanced assimilation method like Kalman filter, 3DVAR, and 4DVAR. We adopted, in the system, a simple nudging method as an assimilation.

We adopted a time retrospective nudging method in which the model variables are gradually tending to observations. This method does not guarantee the optimum state. But it is useful for an operational system as the first step. Though recent developments of computer and soft wares make it possible to adopt advanced assimilation methods, we start a simple system without heavy computation. See, e.g., Anderson (1991), Bennett (1992), Wunsch (1996), Tsuyuki (1997), Cohn (1997), Courtier (1997), Bourtier & Courtier (1999) for the advanced methods.

Seasonal variation in the western North Pacific is important to be investigated with the data assimilation. Realistic reproduction of the past ocean state (reanalysis) is also important. Predictability in the Kuroshio region is not clear, then we examined preliminary predictability experiments. Separation problem and a new assimilation method for each meso–scale eddy are also discussed briefly.

This paper's structure is as follows: We developed an eddy permitting OGCM in the North Pacific. Spinup with climatological forcing fields and analyses of observed data are mentioned in the Section 2. Data assimilation and predictability experiments are mentioned in the Section 3. Section 4 is devoted to results.

2. System, ocean model and data

The system COMPASS–K consists of ocean general circulation model, analyses of observed data and assimilation techniques. Figure 1 shows a block diagram of the system.

The North Pacific Model is an eddy permitting version of the OGCM of the Meteorological Research Institute (MRI). Area for the time integration is from 119°E to 109°W, and 12.5°N to 55.5°N in the North Pacific. Primitive equations of momentum and tracers (temperature and salinity) are solved with finite difference schemes (Bryan, 1969; Kimura and Endoh, 1989). Arakawa's B–grid system is adopted with variable grid sizes: $1/4^{\circ} \times 1/4^{\circ}$ from 23°N to 45°N and 120°E to 180°. The spacing gradually changes to 0.5° in latitude and to 1.5° in longitude outside of the region. The model has 21 vertical levels. Maximum depth is 4500 m. There are 5 levels in the upper 200 m. Used are realistic coastal and bottom topographies without any smoothing. A slope–advective bottom topography scheme

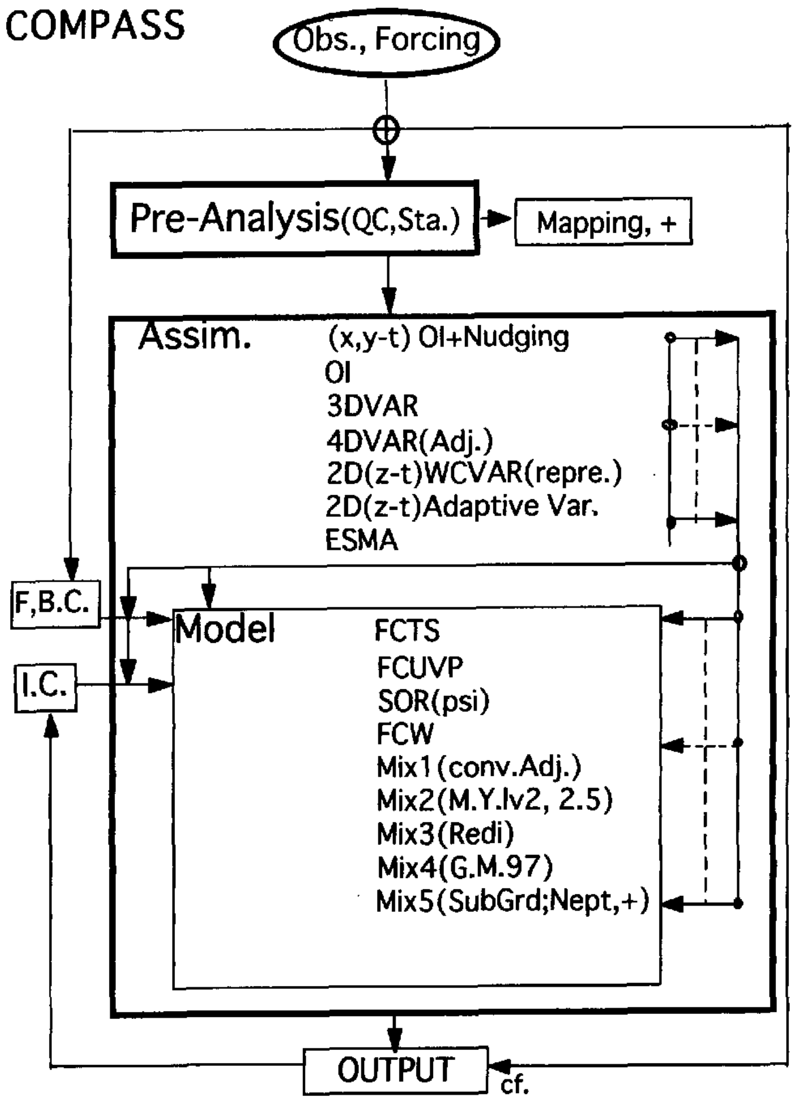


Fig. 1. Block diagram of the assimilation system.

is also adopted for a correct calculation of flows around a steep bottom topography (Ishizaki and Motoi, 1999). Generalized Arakawa scheme is adopted for the momentum advection terms (Arakawa, 1966; Arakawa and Lamb, 1977; Ishizaki and Motoi, 1999). A new definition of the equation of state is also adopted (Ishizaki, 1994). Biharmonic and harmonic types are adopted for dissipation and diffusion terms, respectively. Isopycnal diffusion schemes (Redi, 1982; Gent and McWilliams, 1990) and biharmonic diffusion scheme are also adopted as options. Biharmonic diffusion and dissipation terms are useful in eddy-resolving (or permitting) models, because of selective diffusion / dissipation. An implicit scheme is adopted for a vertical diffusion and dissipation. The typical values of the coefficients of the horizontal viscosity A_H, vertical viscosity A_v, horizontal diffusivity A_{KH}, and vertical diffusivity A_{KV} in the western North Pacific high resolution region are $A_H = 8 \times 10^{10} m^4\ s^{-1}$, $A_V = 1 \times 10^{-5} m^2\ s^{-1}$, $A_{KH} = 250\ m^2\ s^{-1}$, $A_{KV} = 1 \times 10^{-5} m^2\ s^{-1}$. The values of the coefficients are changed according to a function of the space resolution, which functional form is Richardson's 4 / 3–power low (Richardson and Stommel, 1948; Stommel, 1949). Non–slip conditions are adopted at the coast. The model uses a rigid lid approximation. The surface height is, therefore, not the model variable, then a vertical projection from surface height (altimetry) to subsurface temperature and salinity fields is adopted. General frame is the same as the MRI–OGCM reported by Kimura and Endoh (1989).

The model was spun up for 150 years to obtain a statistically equilibrium state. The initial conditions have horizontally uniform profiles of climatological temperature (T) and salinity (S) fields by Levitus (1982) in each model depth, and no–motion. Climatological monthly mean wind stress is adopted as a forcing (Hellerman and Rosenstein, 1983). Time interpolation by Killworth (1996) is adopted. At the northern and southern boundaries and the sea surface, temperature (T) and salinity (S) are restored to the climatological monthly mean values by Levitus (1982). Time constant is 5 days at the surface and 300 days at the bottom, and linearly interpolated between the surface and bottom at the northern and southern boundaries. Though this type of the spinup needs longer time than that started from a climatological T, S distributions, formed are water masses that are consistent with both of boundary conditions and the model state.

Figure 2 shows a horizontal distribution of the barotropic volume transport in January 15th in the 87th year. Contour interval is 10 Sv (Sv = $10^6 m^3\ s^{-1}$), and the shaded region shows cyclonic circulation. Kuroshio volume transport changes about 50–70 Sv. The stream function is positive off Tohoku because of the model Kuroshio's separation. Volume transport in the subpolar gyre is about 10–40 Sv, and it has a seasonal change (minimum in summer). Figure 3 shows a snapshot of the velocity field at 500 m depth. It shows a realistic velocity field.

Model has a realistic salinity distribution with a minimum value (less than 34.4 psu) at 750 m depth in a vertical north–south section along the date line (figure omitted). This model does not adopt the isopycnal diffusion scheme by Redi (1982) and Gent and McWilliams (1990) in this study. Because the model resolves mesoscale eddies, the eddies reproduce the realistic salinity distribution.

TOPEX / POSEIDON altimeter data are analyzed according to Kuragano and Shibata (1997) and Kuragano and Kamachi (2000) with historical ship data. A space–time interpolation method is adopted for obtaining values at the model grids (Kuragano & Kamachi, 2000). Three years, 1993 to 1995, of the TOPEX / POSEIDON altimeter data have been used to produce estimates of the statistical space–time scales of global oceanic variability for the sea surface height. We evaluated distributions of the best–fit decorrelation scales of the second order autoregressive process from obsevations. The decorrelation scales of the TOPEX / POSEIDON (T / P) altimetry data in a space–time domain show inhomogeneous and anisotropic features (Kuragano & Kamachi, 2000). The typical space scale is 170–260 (140–160) km, and time scale is 24–25 (41–51) days in the Kuroshio (extension) region. The JMA Ten–Day Marine Analysis, in which in situ observations are used, shows a distribution with rather smaller scales. The typical decorrelation scales in the Tokara strait are 400

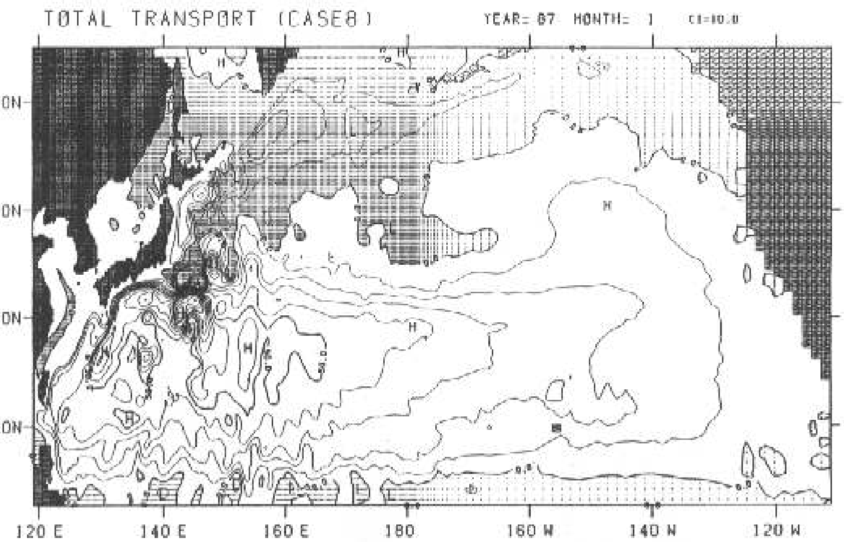


Fig. 2. Stream function of simulation in January, 15th. Contour interval is 10 Sv.

(150) km in 100 (200) m depths. The scales are 170–300 (120) km in 200 (1000) m depths in the Kuroshio south of Japan. They are rather smaller in the Kuroshio extension (10–15 km) and Oyashio (40 km) regions. The scales are 50–80 km in the Japan Sea.

The altimetry has the sea surface information, though it contains the subsurface information about the density structure. When we use a rigid–lid OGCM and an assimilation method such as nudging, we need a statistical evaluation of the subsurface temperature and salinity fields.

Dynamical normal mode decomposition is a candidate for the vertical projection from surface to subsurface. Surface elevation and sea surface temperature (SST) are complimentary with each other. Surface elevation has a larger contribution to the barotropic and lower baroclinic modes like a low–pass filter. Instead the SST has a larger contribution to the higher baroclinic modes like a high–pass filter. In principle, we can define an inverse problem using those complimentary contributions from the surface elevation and SST to the subsurface density fields (see Hsieh, 1985 and Kamachi, 1990). However Woodgate and Killworth (1993) showed that the barotropic and lower baloclinic modes are very sensitive to the vertical density stratification, and the sensitivity is about thousand times of between the surface elevation and the subsurface density structure. Then derivation of the surface pressure (or elevation) from the subsurface density structure or vice versa is not realistic or not practical. On the other hand, methods of statistical inversions have been developed. EOF (De Mey and Robinson, 1987) or a regression method (Mellor and Ezer, 1991) are used for calculating subsurface density fields from the surface elevation.

We adopt a statistical vertical projection method from the altimetry and SST to subsurface temperature and salinity fields by Mellor and Ezer (1991) as follows. We calculate correlations between subsurface density structure (i.e., density: ρ, temperature: T, and salinity

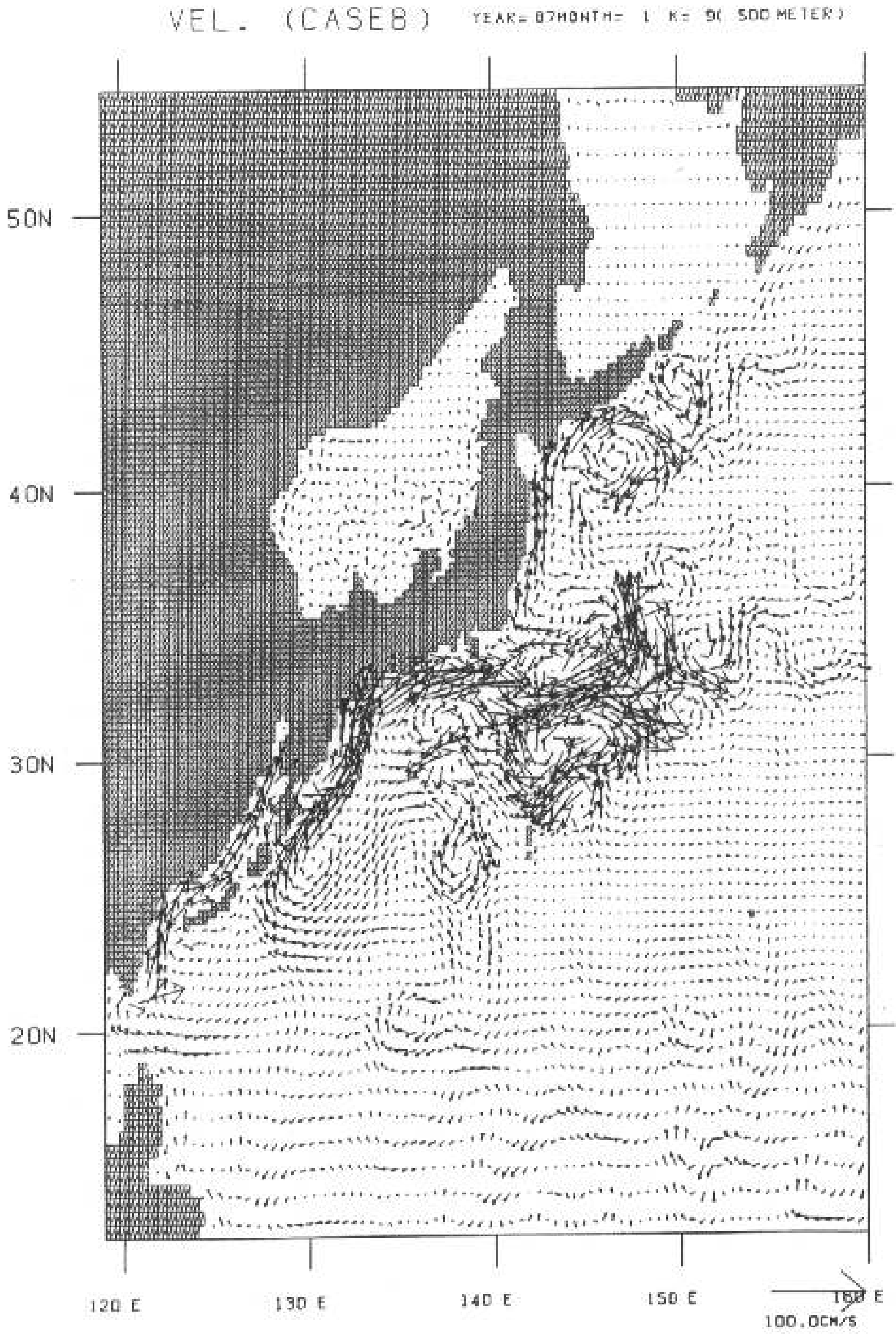
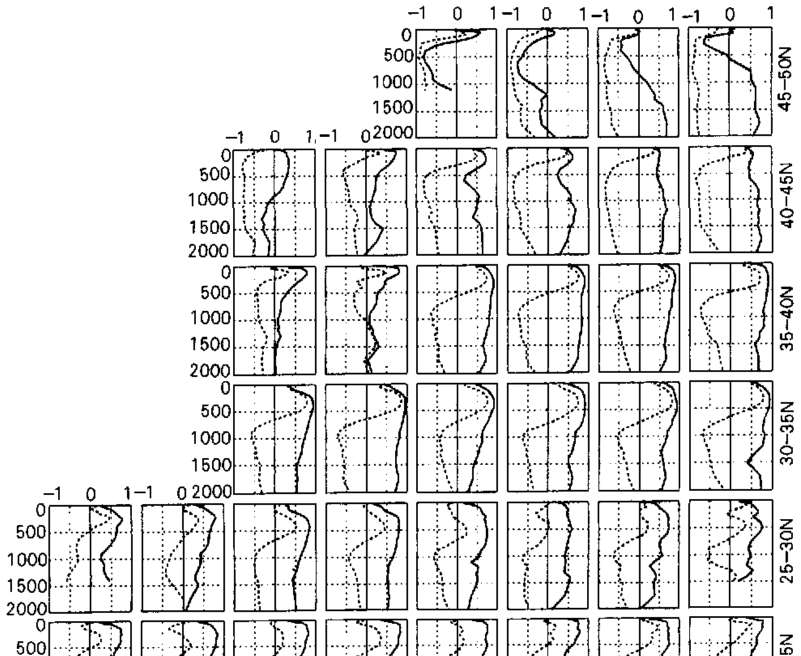


Fig. 3. Velocity field at 500 m depth in January 15th.

S) of historical CTD data and the surface height anomaly $\delta\eta$ of T / P where $\delta(\ )$ shows the difference from climatology. The correlation for temperature is, for example, $C_{T_\eta} = <\delta T \delta\eta> / \{<\delta T^2><\delta\eta^2>\}^{1/2}$, where $<\ >$ represents an ensemble mean in time. The required updates in subsurface water properties can be written $\Delta T = F_{T_\eta} \Delta\eta$, where $\Delta(\ )$ represents the updating quantity before OI, and the correlation factor is $F_{T_\eta} = <\delta T \delta\eta> / <\delta\eta^2>$. The factor is related to a vertical gradient of the density that is proportional to the local potential vorticity.

Mellor and Ezer (1991) adopted a statistics calculated from the model. In our system, we adopted ship observations by the World Ocean Atlas (Levitus and Boyer, 1994; Levitus et al., 1994) for calculating the correlation and regression coefficient (Kuragano and Kamachi, 1997). Figure 4 shows a geographical distribution of the vertical correlation coefficients. Fig. 4a shows the correlation of altimetry versus temperature and salinity. Fig. 4b shows SST versus subsurface temperature and salinity fields. The correlation coefficients of altimetry have the maximum values at about 400 m depth, and have higher values shallower than 800 m depth. On the other hand, SST correlation has a shallower influence depth (about 100 m depth). The complimentary feature is related to the vertical normal modes (Hsieh, 1985; Kamachi, 1995). Though the projected temperature and salinity fields have a realistic distribution (figure is omitted), Kuroshio front is weak and the width is broad south of Japan. The statistics of the vertical projection is calculated every 5 degree. The statistics should reflect



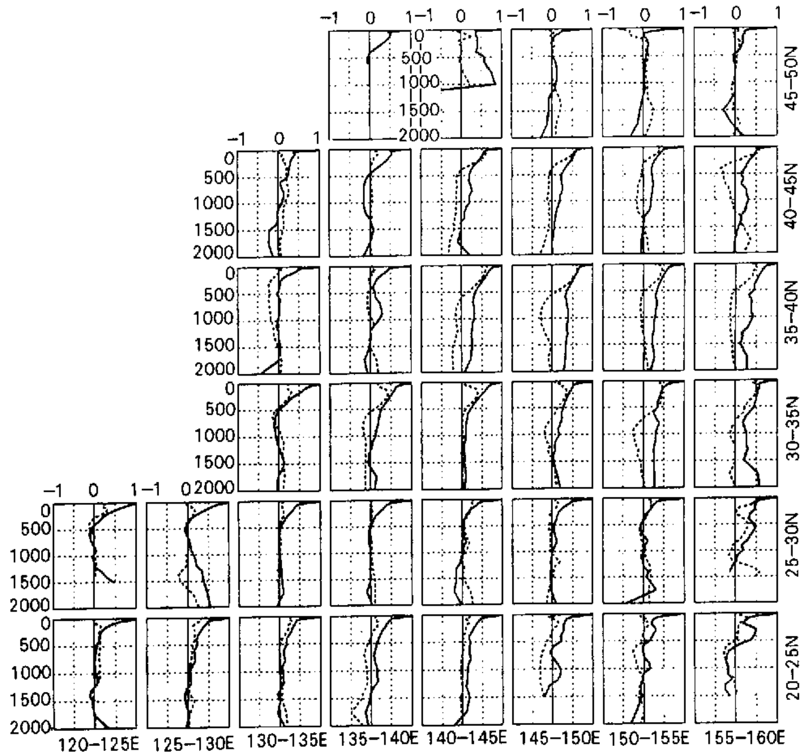


Fig. 4. Geographical distribution of the vertical correlations. (a): Altimetry versus temperature (solid line) and salinity (broken line). (b): SST versus temperature (solid line) and salinity (broken line).

ocean phenomena, then correlation should be calculated in the smaller area which boundary depends on ocean phenomena. Also the correlation is calculated once. It should be calculated in each season. Those recalculations will improve the assimilation results. It is future's study. Calculated 5-day mean values of temperature and salinity are adopted for the assimilation experiments in the study.

3. Assimilation and predictability experiments

The T / P altimeter data are compiled with an optimum interpolation method in a horizontal space and time domain each 5 days. Subsurface T and S fields are evaluated with the correlation scheme of a statistical vertical projection from the surface T / P altimetry. According to the correlation map, we adopt the data shallower than 925 m for assimilation.

A nudging method is adopted for the assimilation. The time scale of the inverse of the coefficient is 5 days in these experiments. The method has a delayed response about the time scale that is related to the inverse of the coefficient of the nudging term. Therefore a time-retrospective scheme was adopted: Observed data are started to be inserted into the model from 3 days before the observation time. Therefore the response started from the day, and the model state reaches closely to the observation state in the observation time. The nudging is stopped at the observation time. One interval of the observation insertion is 5 days, then 2 days are free evolution. After the nudging, the model state remains in the period of the free evolution.

Modeling and assimilation have not reproduced a realistic eddy activity well, which affects the variability of the Kuroshio path and the separation and variability in the confluence area between subtropical and subpolar gyres. An eddy synthetic model approach (ESMA) is developed for reproducing density and velocity fields of mid−ocean eddies into an eddy resolving ocean general circulation model (Uboldi, 1998; Kamachi et al., 1999). Geometric features of temperature and salinity profiles are generated from historical data, and are included in the synthetic model. Density field consists of four types of features for the eddies. Three types of velocity fields are derived: optimum interpolation with another ocean data, gradient velocity and spin−up velocity with the OGCM. Spinup velocity fields are adopted in the COMPASS−K. Then those temperature, salinity and velocity data are combined with ship observation through an optimum interpolation, and are assimilated. One month assimilation and free evolution prediction show a realistic eddy propagation (figure is omitted).

When we adopt an advanced assimilation technique, the velocity is updated / improved by density observation through error covariance matrices and / or constraints. On the other hand, when we adopt a technique such as an optimum interpolation or a nudging, we have several schemes of the velocity updating (e.g., Haines, 1994; Kamachi, 1995; Cooper and Haines, 1996). We adopted no−updating of velocity in these experiments.

Two types of assimilation experiments are carried out: climatological seasonal variation and real assimilation (reanalysis) experiments.

At first we start to show results of climatological experiments. Figure 5 shows a snapshot

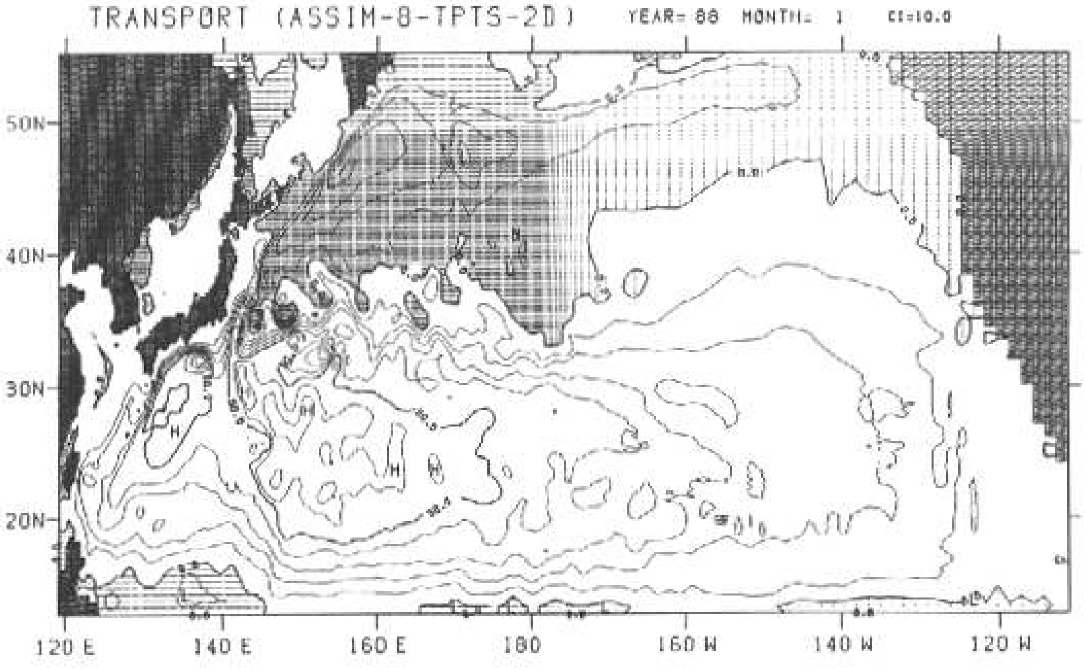


Fig. 5. Total transport after assimilation in January 15th.

of the stream function that means barotropic volume transport in January after assimilation. Comparing to Fig.2, the subpolar gyre has a stronger transport. Volume transport of the Kuroshio is realistic, smaller than the simulation. Comparison with independent data will be reported in later.

Figure 6 shows a velocity field at 500 m depth in January 15th after assimilation. Separation and path of the Kuroshio is reproduced well. Oyashio is also reproduced well. East of Okinawa and Amami islands have a continuous northward flow. The flow is steady in this system, it may be due to weak eddy activities in the climatological experiments, and it is not consistent with some observational evidence. Further investigations are needed from observations and assimilation.

Figure 7 is an example of horizontal distributions of (a) velocity, (b) temperature and (c) salinity after assimilation at 100 m depth in June in an assimilation experiment. The distribution of the velocity, temperature and salinity fields of the Kuroshio are more realistic than simulated fields. The separation point is realistic, while the point in the simulation experiments is far north near Hokkaido (omit). Ship observation compiled in the Japan Meteorological Agency (JMA) shows that the assimilation captures the distributions of temperature and velocity fields well. Figure 8 shows a vertical distribution of salinity in the north–south section along the date line. It shows a realistic distribution and intrusion of salinity minimum.

Figure 9 shows comparisons of time series of month–to–month variation of volume transports by observation, simulation and assimilation of Kuroshio in the Tokara strait, along ASUKA line, in the Oyashio region off Tohoku, and PM–line in the Japan Sea. Figure 9a shows the monthly transport in the Tokara strait calculated from March 1992 to March 1995 by mooring buoys (Misumi et al., 1997). The mean climatological transport is between 200 and 800 m depth. The Kuroshio volume transport in the Tokara strait is not improved by the assimilation. The discrepancy is due that the observation area is between the T / P orbits and the space scale adopted in the analysis is larger. In the area we need ship and float observations for getting realistic ocean state and prediction of the Kuroshio large meander. The Kuroshio volume transport shallower than 1000m depth, in Fig. 9b, shows the similar month–to–month variation to an observation result of ASUKA line off Shikoku (Imawaki et al., 1997). It has not a clear annual variation because of a bottom pressure torque which effect on the weak annual variation in the Kuroshio transport has been clearly explored by Kagimoto and Yamagata (1997). The volume transports of the Oyashio (Uehara et al., 1997) and PM–line in the Japan Sea are also improved, Fig. 9c and 9d. In the Oyashio region, the data show the seasonal variation from 5 to 10 Sv. In these figures, the differences between assimilation and observation are less than 10 to 20 Sv. 10 Sv transport difference is equivalent to a few cm / sec velocity error in the ASUKA line for example. Therefore the results for ASUKA, Oyashio and Japan Sea have good agreements with observations. In the Tokara strait, in which the observation line is between two orbits of T / P, propagation of information of T / P data is not enough in our optimum interpolation method.

Next we carried out three–year assimilation (reanalysis) experiment in 1993 to 1995. Observed data are T / P altimetry and temperature data that are compiled in JMA. Figure 10 shows an example of the velocity field at 10 m depth and temperature field at 100 m depth in September 15th, 1995. Both fields are reproduced well. The volume transports along the ASUKA line are shown in Fig. 11 in 1993 as an example of a comparison of assimilation and

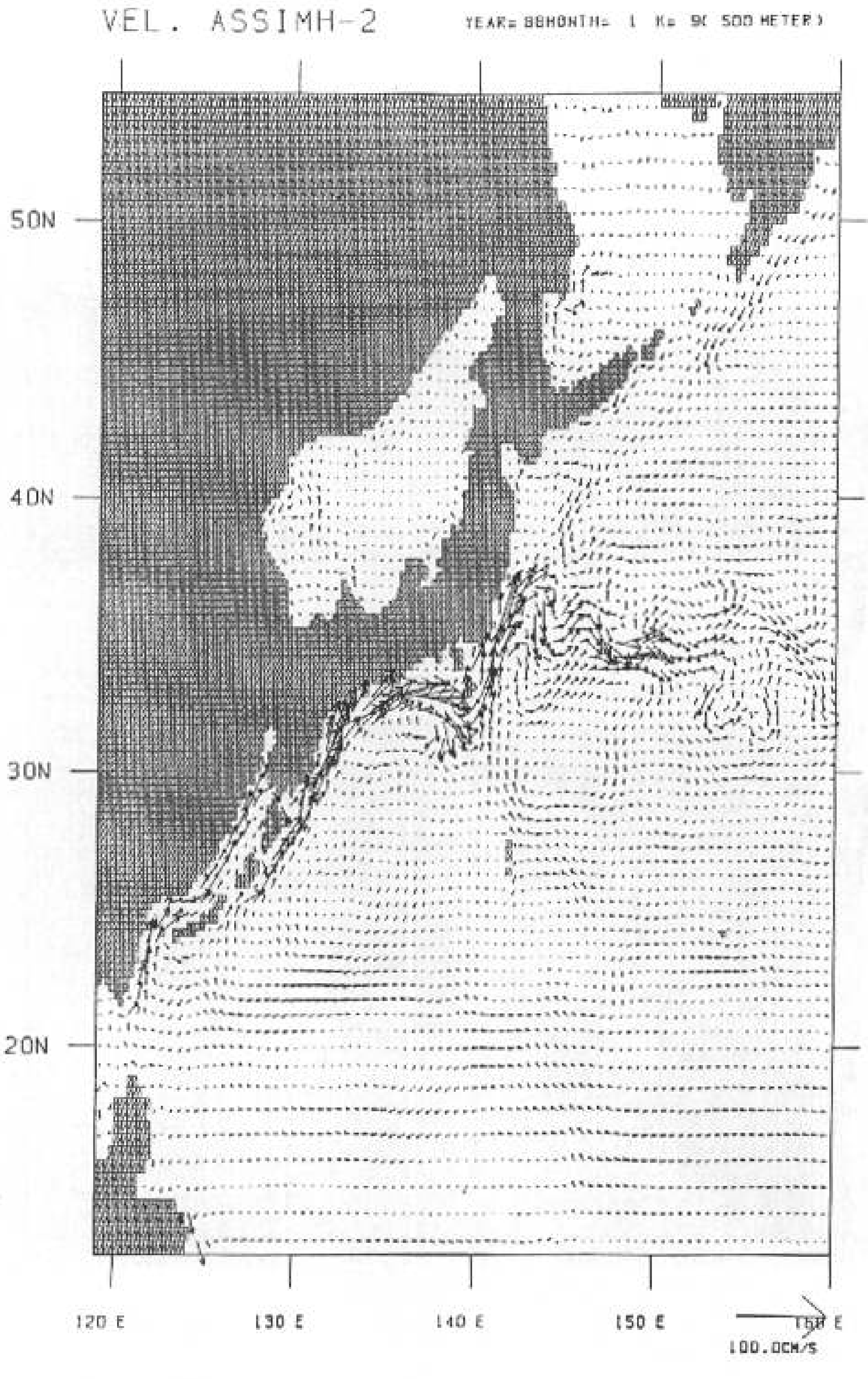
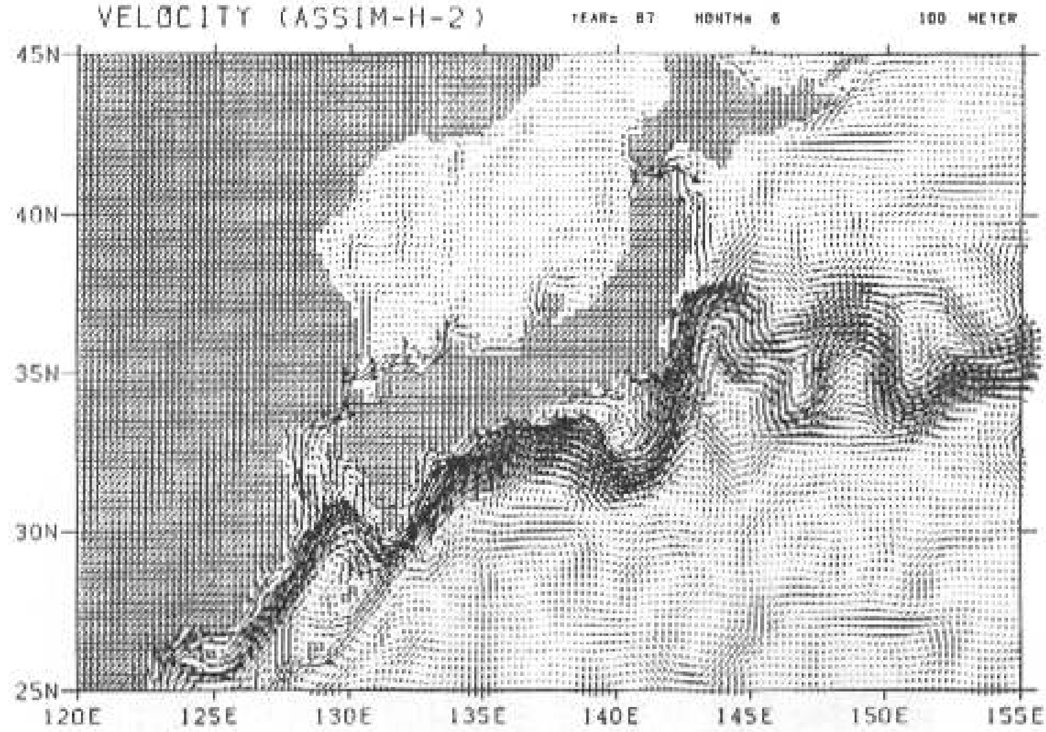
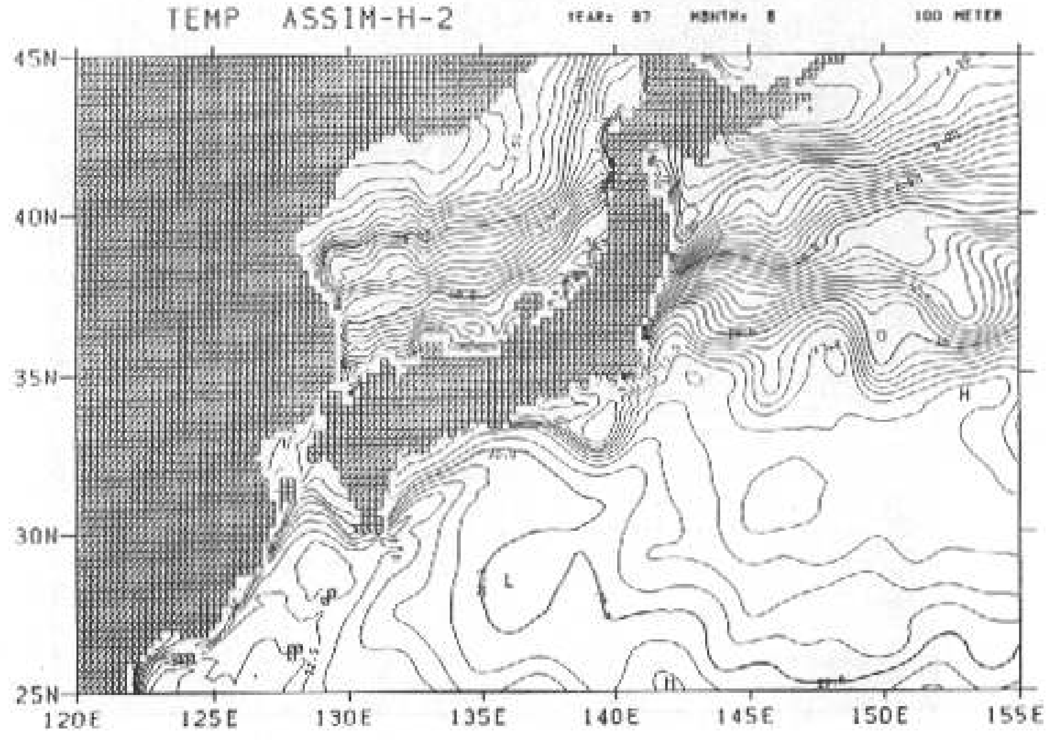


Fig. 6. Velocity field at 500 m depth in January 15th after assimilation.


VELOCITY (ASSIM-H-2)
YEAR= 87 MONTH= 8
100 METER
45N
40N
35N
30N
25N
120E
125E
130E
135E
140E
145E
150E
155E


TEMP ASSIM-H-2
YEAR= 87 MONTH= 8
100 METER
45N
40N
35N
30N
25N
120E
125E
130E
135E
140E
145E
150E
155E

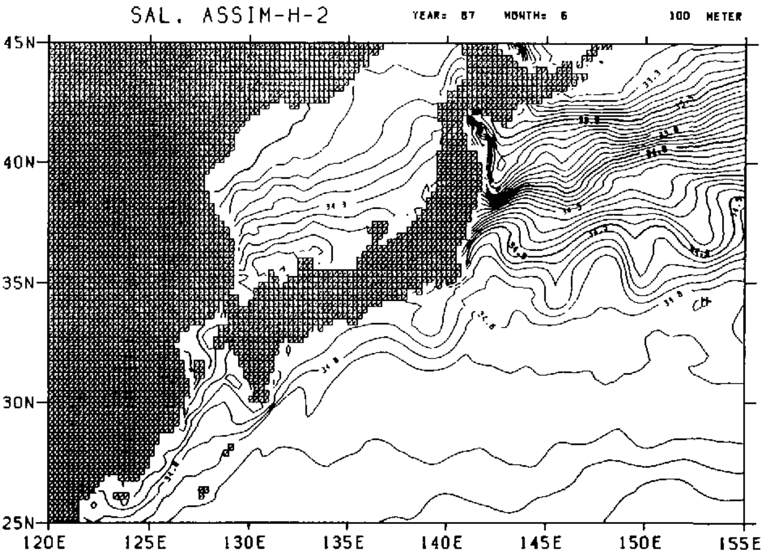


Fig. 7. Horizontal distribution of (a) velocity, (b) temperature and (c) salinity after assimilation at 100 m depth. Contour intervals are 0.1°C and 0.1 psu for temperature and salinity, respectively.

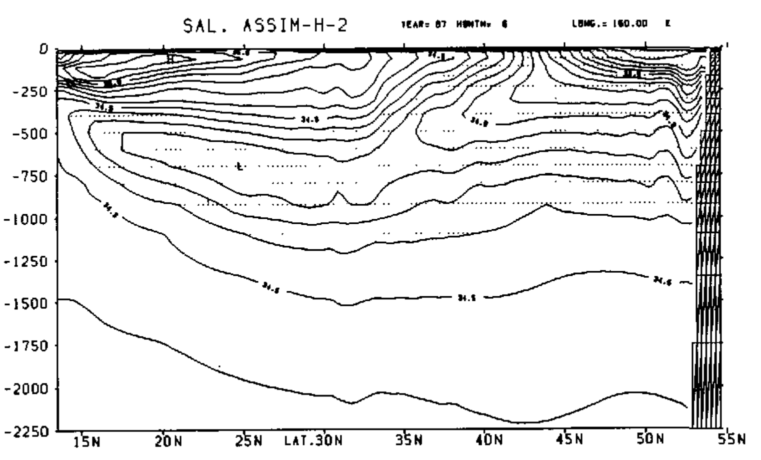
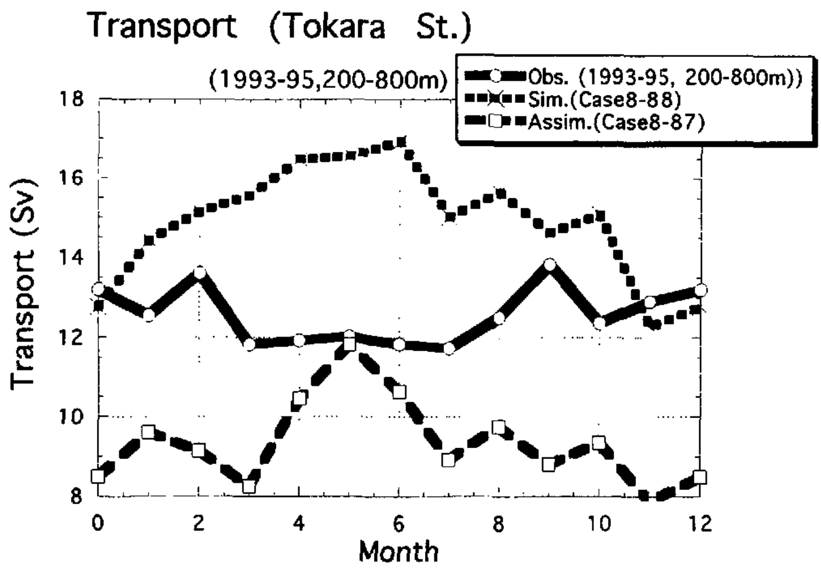
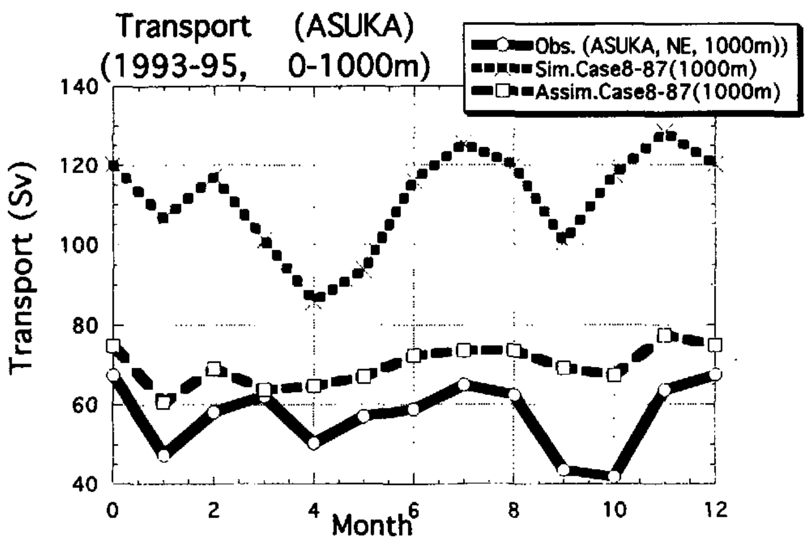
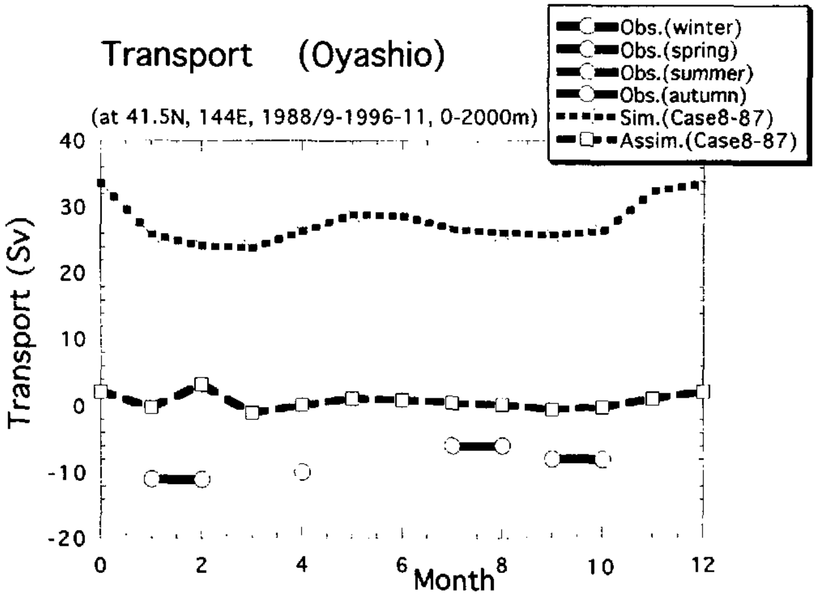


Fig. 8. Vertical distribution of salinity in the north–south section along the date line. Contour interval is 0.1psu. Shaded region means that salinity is lower than 34.4 psu.


Transport (Tokara St.)
(1993-95,200-800m)
Obs. (1993-95, 200-800m))
Sim.(Case8-88)
Assim.(Case8-87)
Transport (Sv)
Month
18
16
14
12
10
8
0
2
4
6
8
10
12


Transport (ASUKA)
(1993-95, 0-1000m)
Obs. (ASUKA, NE, 1000m))
Sim.Case8-87(1000m)
Assim.Case8-87(1000m)
Transport (Sv)
Month
140
120
100
80
60
40
0
2
4
6
8
10
12



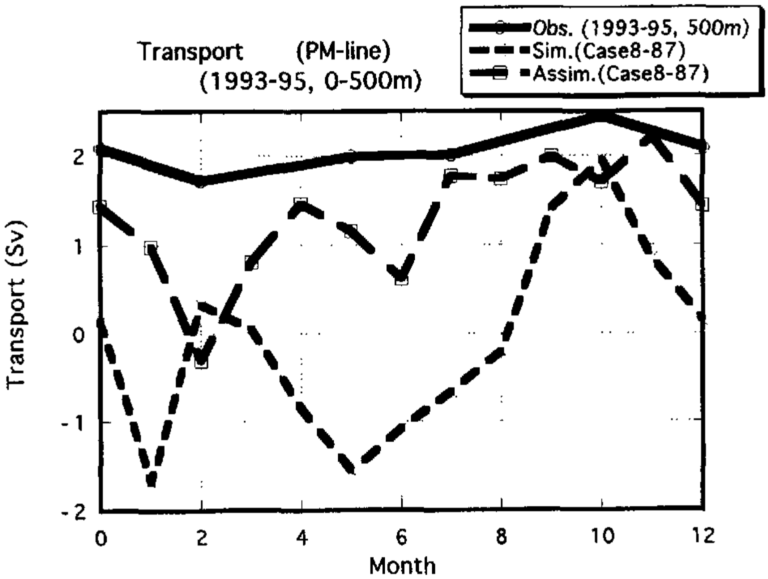


Fig. 9. Comparisons of time series of month–to–month variation of volume transports by observation, simulation and assimilation of Kuroshio and Oyashio (a) in the Tokara atrait, (b) along ASUKA line (shallower than 1000 m north of 30°N), (c) in the Oyashio region off Tohoku (between 1000 m and 2000 m depths, 28 km east–west width around 41.5°N, 144°E), and (d) PM–line in the Japan Sea. Solid lines show observations, broken lines show assimilated results, and dotted lines show simulation results.

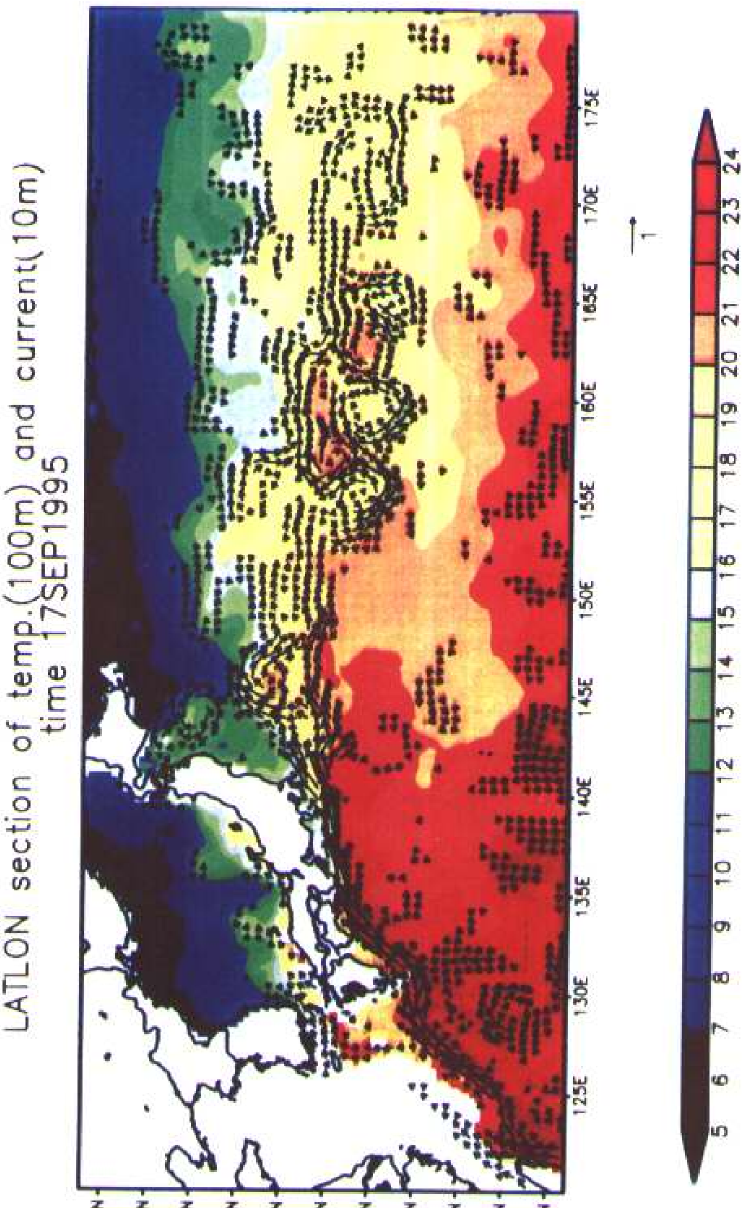


Fig. 10. Horizontal distribution of velocity at 10 m depth and temperature at 100 m depth in September 15th, 1995.

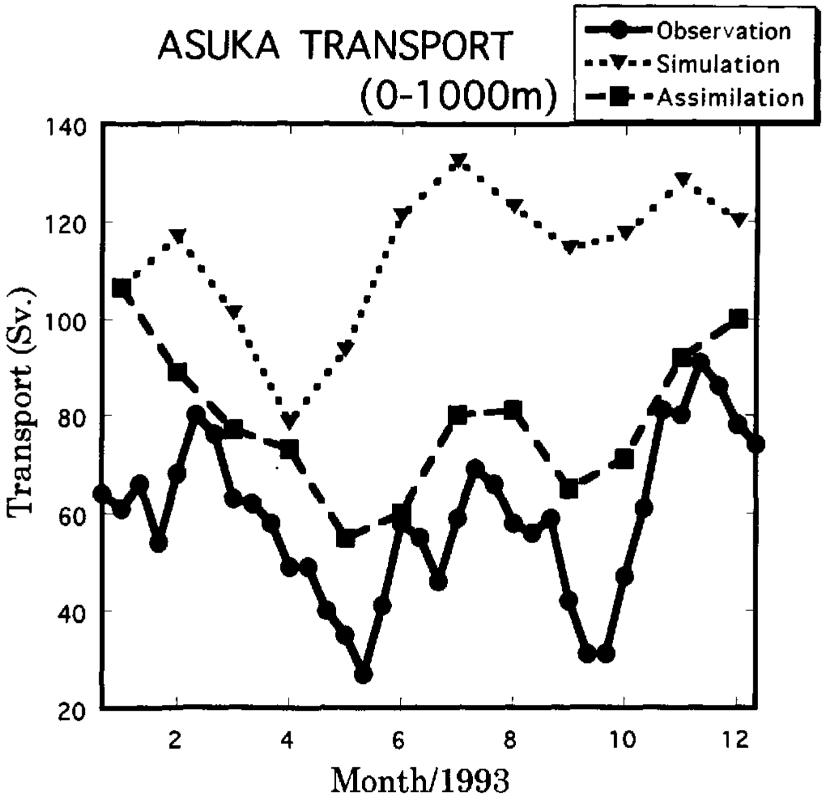


Fig. 11. Time series of month–to–month variation of volume transport along the ASUKA line in 1993.

independent observation. Variability of the transport is captured well. We also tried to examine the predictability of the ocean state around Japan. Prediction run is started from February 15th, 1994. The time–growth curve of the forecast error can show a time scale of predictability.

In the following, the outline of a predictability experiment is described briefly. Assimilated initial conditions (ICs) are obtained once a day from February 15th to March 14th ($i=1,\cdots, N$; $N=28$ is the number of the initial conditions and ensemble members). From each IC, forecast runs are done. Forecasted ocean states are restored once a day with the lead time j or k (j and $k=0,\cdots, M$; $M=27$ is the number of the lead time; unit is day). We adopted temperature value at 100 m depth along the ASUKA line as the predictability variable. According to the method detailed by Lorenz (1982), the statistics is defined by

$$E_{jk}^2=\frac{1}{N}\sum_{i=1}^{N}\frac{1}{L_x L_y}\iint_{ASUKA}[T_{ij}(x,y,z=100\ m)-T_{ik}(x,y,z=100\ m)]^2\,dxdy, \tag{1}$$

where $T_{ij}(T_{ik})$ is a predicted temperature data started from ith day with the lead time j day

($k$day). E is a statistical value of the root mean square difference of the forecast error with different lead times (j and k). E is plotted against k in the predictability diagram Fig. 12, the lower bound is estimated from the thick curve of $0-k$ in Fig. 12. The smallest initial error is 0.1°C at 1st–day lead time ($k=1$). It has doubled with the 2nd day lead time. Thus the doubling time is 1 to 2 days. It is much smaller than the former studies by Carton (1987), Adamec (1989) and Brasseur et al. (1996). If we adopt the 3–day time step such as Brasseur et al. (1996), instead of 1–day, we get 10 days for the doubling time scale. On the other hand, the lowest curve ($k-j=1$) in Fig. 12 represents the smallest errors in all ensemble members. The minimum value is about 0.033°C at 9th day ($k=9$). The doubling of the error is given at 26th day ($k=26$). Then the doubling time is about 17 days. If we adopt the 3–day time step, we get the minimum value at 11th day on the curve $k-j=3$, and the doubling value is given at 20th day. The time scale is 10 days. The time scales in this study, which are for predicting a local (ASUKA line) subsurface temperature, are much smaller than the former studies by Auer (1987), Mellor and Ezer (1991) and Brasseur et al. (1996) in the Gulf Stream region. It depends on the different predicted variabilities, model, observed data, time step size for the ensemble members, area (global or local) and the method. More comprehensive examination with the different conditions will be done in a future study.

4. Results

We introduced an ocean data assimilation system and results of analyses of T / P altimetry, assimilation and predictability experiments. The model can reproduce the realistic

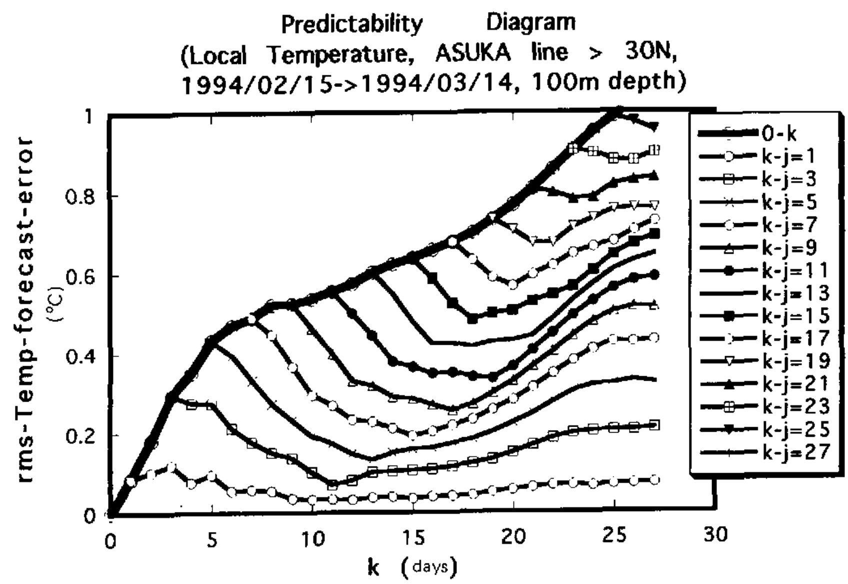


Fig. 12. Predictability diagram of temperature along the ASUKA line (north of 30°N) from February 19th, 1994.

ocean state. Assimilated results show good reproduction of volume transport, temperature and velocity fields, though some areas are difficult for improvement. Additional observation (ship and float) or advanced assimilation methods such as adjoint method and adaptive variational method (Zhu and Kamachi, 2000) may be able to solve the problem. Time scale of the predictability is 17 days that is much smaller than the former studies in the Gulf Stream region.

Recent developments of the data assimilation are adopted in operational centers. Data assimilation is not only an interpolation / extrapolation of observed data but a total system for now– / fore–casting and understanding of oceanic & climate phenomena. We need to continue to develop each components of the system. The system gives the information of heat content of the ocean, heat transport and grasp of its variability for climate research. Our system is also a prototype for the GODAE (Global Ocean Data Assimilation Experiment).

A part of this work is supported by CREST (Core Research for Evolutional Science and Technology) of Japan Science and Technology Corporation (JST).

REFERENCES

Adamec, D., 1989: Predictability of a quasi–geostrophic ocean flow; Sensitivity of varying model vertical resolution. *J. Phys. Oceanogr.*, **19**, 1753–1764.

Anderson, D.L.T., 1991: Data assimilation in ocean models. In: M. Latif (Ed.), Strategies for Future Climate Research. *Max–Plank–Institute fuer Meteorologie*, Hamburg, Germany, 193–225.

Arakawa, A., 1966: Computational design for long–term numerical integration of the equations of fluid motion: Two–dimensional incompressible flow, Part. I. *J. Comp. Phys.*, **1**, 119–143.

Arakawa, A., and V. R. Lamb, 1977: Computational design of the basic processes of the UCLA general circulation model. *Methods in Comp. Phys.*, **17**, 173–265.

Auer, S.J., 1987: Five–year climatological survey of the Gulf Stream system and its associated rings. *J. Geophys. Res.*, **92, C11**, 11709–11726.

Bennett, A.F., 1992: *Inverse Methods in Physical Oceanography*. Cambridge University Press, Cambridge, 346 pp.

Bourtier, F., and P. Courtier, 1999: *Data assimilation concepts and methods*. Rep. of ECMWF, 75 pp.

Brasseur, P., E. Blayo, and J. Verron, 1996: Predictability experiments in the North Atlantic Ocean: Outcome of a quasi–geostrophic model with assimilation of TOPEX / POSEIDON altimeter data. *J. Geophys. Res.*, **101**, 14161–14173.

Bryan, K., 1969: A numerical method for the study of the world ocean. *J. Comp. Phys.*, **4**, 347–376.

Carton, J. A., 1987: How predictable are the geostrophic currents in the recirculation zone of the north Atlantic? *J. Phys. Ocenogr.*, **17**, 751–762.

Cohn, S.E., 1997: An introduction to estimation theory. *J. Met. Soc., Japan.*, **75**, 257–288.

Cooper, M., and K. Haines, 1996: Altimetric assimilation with water property conservation. *J. Geophys. Res.*, **101**, 1059–1077.

Courtier, P., 1997: Dual formulation of four–dimensional variational assimilation. *Q. J. Roy. Met. Soc.*, **123**, 2449–2461.

De Mey, P., and A. Robinson, 1987: Assimilation of altimeter eddy fields in a limited–area quasi–geostrophic model. *J. Phys. Oceanogr.*, **17**, 2280–2293.

Gent, P. R., and J. C. McWilliams, 1990: Isopycnal mixing in ocean general circulation model. *J. Phys. Oceanogr.*, **20**, 150–155.

Haines K., 1994: Dynamics and data assimilation in oceanography. in Data Assimilation: Tools for Modelling the Ocean in a Global Change Perspective, ed. by P.P. Brasseur and J.C.C. Nihoul,. *NATO ASI Series I*, **19**, 1–32.

Hellerman, S., and M. Rosenstein, 1983: Normal monthly wind stress over the world ocean with error estimates. *J. Phys. Oceanogr.*, **13**, 1093–1104.

Hsieh, W., 1985: Modal bias in sea level and sea surface temperature, with application to remote sensing. *J. Phys. Oceanogr.*, **15**, 351–356.

Imawaki, S., H. Uchida, H. Ichikawa, M. Fukasawa, S. Umatani, and ASUKA Group, 1997: Time series of the Kuroshio transport derived from field observations and altimetry data. *Int. WOCE Newsletter*, **25**, 15–18.

Ishizaki, H., 1994: A simulation of the abyssal circulation in the North Pacific Ocean. Part I: Flow field and comparison with observations. *J. Phys. Oceanogr.*, **24**, 1941–1954.

Ishizaki, H., and T. Motoi, 1999: Reevaluation of the Takano–Oonishi scheme for momentum advection on bottom relief in ocean models. *J. of Atmos. Ocean. Tech.*, **16**, 1994–2010.

Kagimoto, T., and T. Yamagata, 1997: Seasonal transport variations of the Kuroshio: An OGCM simulation, *J. Phys. Oceanogr.*, **27**, 403–418.

Kamachi, M., 1995: Toward assimilation of altimetry in combination with conventional data in the North Pacific Ocean. *Proceeding of Int. Workshop on Num. Pred. of Ocean. Cond.*, Tokyo, 59–66.

Kamachi, M., T. Kuragano, T. Yoshida, F. Uboldi, and N. Yoshioka, 1998: An Ocean Data Assimilation System. *Weather Service Bulletin, JMA*, **65**, 1–19 (in Japanese).

Kamachi, M., T. Kuragano, and D. F. Uboldi, 1999: A reproduction of Kuroshio separation and data assimilation of KWCR in the confluence area. *In "Ecosystem Dynamics of the Kuroshio–Oyashio Transition Region"*, Japan Marine Science Foundation, 9–23.

Killworth, P.D., 1996: Time interpolation of forcing fields in ocean models. *J. Phys. Oceanogr.*, **26**, 136–143.

Kimura Y., and M. Endoh, 1989: Response experiment of the Pacific Ocean to anomalous wind stress with ocean general circulation model, *Tech. Rep. of the Met. Res. Inst.*, **24**, 96 pp.

Kuragano, T., A. Shibata, 1997: Sea surface dynamic height of the Pacific Ocean derived from TOPEX / POSEIDON altimeter data: calculation method and accuracy. *J. of Oceanogr.*, **53**, 585–599.

Kuragano, T. and M. Kamachi, 1997: Investigation of temperature and salinity fields statistically derived from TOPEX / POSEIDON altimeter data. *Proceedings of the Symposium Monitoring the Oceans in the 2000s: An integrated approach.*, Biarritz, France, 4–29.

Kuragano, T. and M. Kamachi, 2000: The global statistical space–time scales of oceanic variability estimated from the TOPEX / POSEIDON altimeter. *J. Geophys. Res.*, **105**, 955–974.

Levitus, S., 1982: Climatological atlas of the world ocean. NOAA Prof. Paper 13, 174 pp.

Levitus, S. And T.P. Boyer, 1994: World Ocean Atlas 1994, 4: Temperature. *NOAA Atlas NESDIS* 4, 117 pp.

Levitus, S., R. Gurgett, and T. P. Boyer, 1994: World Ocean Atlas 1994, 3: Salinity. *NOAA Atlas NESDIS* 3, 99 pp.

Lorenz, E.N., 1982: Atmospheric predictability experiments with a large numerical model. *Tellus*, **34**, 505–513.

Mellor, G.L., and T. Ezer, 1991: A Gulf stream model and an altimetry assimilation scheme. *J. Geophys. Res.*, **96**, 8779–8795.

Misumi, A., H. Yamamoto, H. Yoshikawa, K. Ishikawa, N. Kanno, N. Yoshioka, H. Kinosita, H. Yoritaka, N. Shikama, T. Yamashiro, M. Sakurai, and A. Maeda, 1997: Report of Kuroshio variability in the Tokara strait. *Proc. of the 1997 spring Mtg. Of the Jpn Ocn. Soc.*, No. 118 (in Japanese).

Redi, M.H., 1982: Oceanic isopycnal mixing by coordinate rotation. *J. Phys. Oceanogr.*, **12**, 1154–1158.

Richardson, L.F., and H. Stommel, 1948: Note on eddy diffusion in the sea. *J. Meteorol.*, **5**, 238–240.

Stommel, H., 1949: Horizontal diffusion due to oceanic turbulence. *J. Mar. Res.*, **8**, 199–225.

Tsuyuki, T., 1997: Variational data assimilation. *JMA–NPD Rep., special volume*, **43**, 102–165 (in Japanese).

Uboldi, F., 1998: Report on Eddy Synthetic Models, Reports in COMPASS Group No.2. *Met. Res. Inst.*, 49 pp.

Uehara, K., H. Miyake, and T. Iwao, 1997: Geostrophic volume transport of Oyashio and Tsugaru strait calculated from CTD observations along 41°30′N. *Proceeding of the 1997 spring Mtg. Of the Jpn Ocn. Soc.*, No. 133 (in Japanese).

Woodgate, R.A., and P.D. Killworth, 1993: Can we derive pressure from density using normal modes? *Ocean Modelling*, **100**, 5–6.

Wunsch, C., 1996: *The Ocean Circulation Inverse Problem.* Cambridge University Press, Cambridge, 442 pp.

Zhu, J., and M. Kamachi, 2000: An adaptive variational method for data assimilation with imperfect models. *Tellus*, **52A**, 265–279.