CHAPTER 2

THEORY

A single level shallow water equation is applied to investigate characteristic of active and break monsoon. The concept of the numerical model are discussed in Section 2.1 while the detail of numerical techniques are explained in Section 2.2. In order to adjust the relationship between wind field and geopotential height to satisfy the assumption in the single level shallow water equation, an initialization method is required as explained in Section 2.3. The details of ensemble forecast techniques are presented in Section 2.4 focusing on the singular vector. Active and break summer monsoon are explained in Section 2.5.

2.1 The Shallow Water Model

The shallow water equation consists of a set of three equations describing the thin layer of constant density fluid or the atmosphere in which the horizontal scale of the flow is much greater than the layer depth. In this research, an incompressible fluid flow (constant density) has been applied in responding to gravitational and rotational accelerations. Simple conditions are used to explain the motions such as an inviscid fluid layer in a rotating frame (no vertical wind shear) and the vertical advection term does not appear in the shallow water equation.

The set of simplification of the shallow water equations can be written in form of the momentum equations (Eq. 2.1 and Eq. 2.2), hydrostatic equation (Eq. 2.3) and continuity equation (Eq. 2.4) (Holton, 2004).

$$\frac{du}{dt} - fv + \frac{1}{\rho}\frac{\partial p}{\partial x} = 0 \tag{2.1}$$

$$\frac{dv}{dt} + fu + \frac{1}{\rho}\frac{\partial p}{\partial y} = 0 \tag{2.2}$$

$$\frac{\partial p}{\partial z} + \rho g = 0 \tag{2.3}$$

$$\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}+\frac{\partial w}{\partial z}=0 \tag{2.4}$$

The total time derivative is given by

$$\frac{d}{dt}=\frac{\partial u}{\partial t}+u\frac{\partial}{\partial x}+v\frac{\partial}{\partial y}$$

where u and v are the horizontal winds components along x and y axes, respectively

w is the vertical wind component along z axis

ρ is the density of the fluid

p is the pressure

g is the acceleration due to gravity

f is the Coriolis parameter which is given by $f=2\Omega\sin\theta$

Ω is the angular velocity of the earth

θ is the latitude.

The system can be simplified by eliminating the vertical velocity w. So the vertical velocity can be neglected with respect to the horizontal winds, thus the z component of the momentum equation will be reduced to the hydrostatic equation (Eq. 2.3).

Consider a thin fluid layer above a flat surface neglecting bottom topography. The depth of the fluid layer is a function of x and y $(h=h(x,y))$ where a mean depth H is defined by

$$H=\iint h(x,y)dxdy. \tag{2.5}$$

Assume that the pressure at the top of the fluid layer is a constant p_0. Integrating the hydrostatic equation (Eq. 2.3) between the limits z and h to get

$$p=\rho g(h-z)+p_0. \tag{2.6}$$

Assume that the pressure at a point given by the weight of fluid layer above it (plus p_0). It implies that the horizontal pressure gradient at a depth z is given by

$$\frac{1}{\rho}\frac{\partial p}{\partial x} = g\frac{\partial h}{\partial x} \tag{2.7}$$

$$\frac{1}{\rho}\frac{\partial p}{\partial y} = g\frac{\partial h}{\partial y}. \tag{2.8}$$

The expressions on the right hand sides are independent of the depth z.

Suppose that the horizontal velocity (u,v) is independent of depth z. The momentum equation shows that a changing in u and v arise through the Coriolis and pressure gradient forces. But both of these are independent of depth z and therefore the accelerations do not vary with z as the velocity will remain independent of depth for all time. This permits a great simplification in which it can be assumed that the velocity (u,v) is constant throughout the fluid layer. Then integrating the continuity equation (Eq. 2.4) with the full depth of the fluid. Since u and v are constant with z the first two terms are given by

$$\int_0^h \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) dz = h\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = h\nabla_H \cdot \mathbf{V} \tag{2.9}$$

Consider the flat plane, the vertical velocity must be vanished. Moreover, the vertical velocity of a fluid particle at the top surface is given by $w(h) = \dfrac{dh}{dt}$. Thus the third term of Eq. 2.4 can be integrated as

$$\int_0^h \left(\frac{\partial w}{\partial z}\right) dz = w(h) - w(0) = \frac{dh}{dt}. \tag{2.10}$$

Finally, the result of integrating the continuity equation (Eq. 2.4) may be written as

$$\frac{dh}{dt} + h\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = 0. \tag{2.11}$$

The set of shallow water equations are

$$\frac{du}{dt} - fv + g\frac{\partial \phi}{\partial x} = 0 \tag{2.12}$$

$$\frac{dv}{dt} + fu + g\frac{\partial \phi}{\partial y} = 0 \tag{2.13}$$

$$\frac{d\phi}{dt} + \phi\nabla_H \cdot \mathbf{V} = 0 \tag{2.14}$$

where ϕ is the geopotential height and is defined as $\phi = gh$.

2.2 Numerical Techniques

2.2.1 Time Integration Scheme

The Matsuno backward time difference scheme is a modified form of the Euler backward time difference scheme. It consists of a trial forward step followed by a backward step giving rise to the called predictor-corrector technique. For the prediction equation (Krishnamutri, 1986)

$$\frac{\partial \zeta}{\partial t} = -J(\psi, \zeta) \tag{2.15}$$

where J is Jacobian operator , ζ is relative vorticity and ψ is stream function (see Eq.2.38).

The trial forward step at the (n+1) time step is given as

$$\left(\zeta_{i,j}\right)_{n+1}^{(1)} = \left(\zeta_{i,j}\right)_n - \left(J(\psi,\zeta)_{i,j}\right)_n \Delta t \tag{2.16}$$

where the superscript (1) denotes the solution of first guess. The forward time step can be assumed as trial solutions of ζ to obtain at all grid points while the backward step is

$$\left(\zeta_{i,j}\right)_{n+1}^{(2)} = \left(\zeta_{i,j}\right)_n + \left(J(\psi,\zeta)_{i,j}\right)_{n+1}^{(1)} \Delta t \tag{2.17}$$

2.2.2 Semi – Lagrangain Advection Scheme

The semi-Lagrangian advection was proposed by Krishnamurtri (1986) and Mathur (1970) and used in the model. If *A*, *B* and *C* represent the forcing functions of the zonal and meridional momentum equations then the continuity equation can be written as

$$\frac{du}{dt} = -g\frac{\partial \phi}{\partial} + fv = A \tag{2.18}$$

$$\frac{dv}{dt} = -g\frac{\partial \phi}{\partial} - fu = B \tag{2.19}$$

$$\frac{d\phi}{dt} = -\phi\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = C \tag{2.20}$$

The semi-Lagrangian advection is called out as follows. As shown in Figure 2.1, parcels located some point *P* at time *t* arrived to a grid point *Q* at time *t*. If *x* and *y* represent the zonal and meridional the distances of *P* from Q then on using Taylor's expansion,

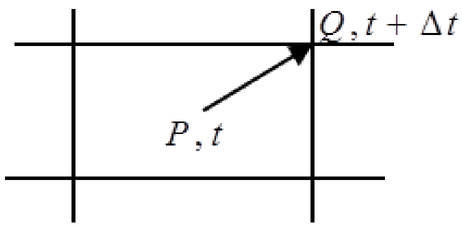

Figure 2.1 Semi-Langrangian advection (Krishnamurtri, 1986).

First guess:

$$x = -u\big|_{Q_t} \Delta t - \frac{1}{2} A\Big|_{Q_t} \Delta t^2 \tag{2.21}$$

$$y = -v\big|_{Q_t} \Delta t - \frac{1}{2} B\Big|_{Q_t} \Delta t^2 \tag{2.22}$$

where the superscript (1) denotes all values of the first guess with the first guess position of *P*. The first values of u, v, A and B at *P* are determined using a 9-point Lagrange interpolation scheme. *F* represents the function to be interpolated, defined as

$$F(P) = \sum_{\substack{i=I-1 \\ j=j-1}}^{\substack{j=J+1 \\ i=I+1}} W_{ij} F_{ij} \tag{2.23}$$

where W_{ij} is the weight for the Lagrange interpolation scheme defined as follows.

$$W_{ij} = \prod_{\substack{k=I-1 \\ k \neq i}}^{k=I+1} \frac{x - x_k}{x_i - x_k} \prod_{\substack{l=J-1 \\ l \neq j}}^{l=J+1} \frac{y - y_l}{y_j - y_l} \tag{2.24}$$

Using the first guess values of $u, v,$ A and B at point P at time t, the second guess position of P is obtained as follows.

Second guess:

$$x = u\big|_{P_t} \Delta t - \frac{1}{2} A\Big|_{P_t} \Delta t^2 \tag{2.25}$$

$$y = v\big|_{P_t} \Delta t - \frac{1}{2} B\Big|_{P_t} \Delta t^2 \tag{2.26}$$

At the new position of P the values of u, v, ϕ, A, B and C at P at time t are formed. The u, v, ϕ values at grid point Q at time $t + \Delta t$ are computed using the predictor – corrector method of the Matsuno time integration scheme. The predictor step consists of a trial forward step and given as (Krishnamurtri, 1986).

Predictor step:

$$u_{Q_{t+\Delta t}} = U\big|_{P_t} + A\big|_{P_t} \Delta t \tag{2.27}$$

$$v_{Q_{t+\Delta t}} = V\big|_{P_t} + B\big|_{P_t} \Delta t \tag{2.28}$$

$$\phi_{Q_{t+\Delta t}} = \Phi\big|_{P_t} + C\big|_{P_t} \Delta t \tag{2.29}$$

The corrector step consists of a backward time differencing step and may be written as follows.

Corrector step:

$$u_{Q_{t+\Delta t}} = U\big|_{P_t} + A\big|_{Q_{t+\Delta t}} \Delta t \tag{2.30}$$

$$v_{Q_{t+\Delta t}} = V\big|_{P_t} + B\big|_{Q_{t+\Delta t}} \Delta t \tag{2.31}$$

$$\phi_{Q_{t+\Delta t}} = \Phi\big|_{P_t} + C\big|_{Q_{t+\Delta t}} \Delta t \tag{2.32}$$

2.3 Initialization

In this research the perturbed conditions are generated from wind only. These conditions will create unbalance between winds field and geopotential height in the linear version of the single level primitive equation (SILEPE) model (Section 3.4). In order to balance the wind and geopotential height fields, the initialization method is done for SILEPE. The governing equations have the assumptions of inviscid, homogeneous, incompressible fluid. Thus, the horizontal equations of motion governing the non-divergent barotropic flow can be written as

$$\frac{\partial u}{\partial t}+u\frac{\partial u}{\partial x}+v\frac{\partial u}{\partial y}=-g\frac{\partial \phi}{\partial x}+fv \tag{2.33}$$

$$\frac{\partial v}{\partial t}+u\frac{\partial v}{\partial x}+v\frac{\partial v}{\partial y}=-g\frac{\partial \phi}{\partial y}-fu \tag{2.34}$$

The continuity equation for the system is

$$\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}=0 \tag{2.35}$$

Given a velocity field V,

where V_ψ is the non-divergent part and V_χ is the divergent part.

$$V=V_\psi+V_\chi \tag{2.36}$$

This implies $\nabla\cdot V_\psi=0$ and $\nabla\times V_\chi=0$. Differentiate Eq. (2.33) by $\frac{\partial}{\partial x}$ and Eq. (2.34) by $\frac{\partial}{\partial y}$ and take the difference between the two equations, the vorticity equation for the non-divergent barotropic flow is obtained.

$$\frac{\partial \zeta}{\partial t}+u\frac{\partial \zeta}{\partial x}+v\frac{\partial \zeta}{\partial y}+v\frac{\partial f}{\partial x}=0 \tag{2.37}$$

where $\zeta=\frac{\partial v}{\partial x}-\frac{\partial u}{\partial y}$ is the relative vorticity. The non-divergent of flow can be expressed in terms of streamfunction, ψ, such that

$$u_\psi=-\frac{\partial \psi}{\partial y} \text{ and } v_\psi=-\frac{\partial \psi}{\partial x}$$

In the streamfunction form, the relative vorticity, ζ, may be expressed as follows

$$\zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} = \nabla^2 \psi \tag{2.38}$$

The non-divergent barotropic vorticity equation Eq. (2.36) can be written as

$$\frac{\partial}{\partial t} \nabla^2 \psi = -J(\psi, \nabla^2 \psi) - \beta \frac{\partial \psi}{\partial x} \tag{2.39}$$

where $-J(\psi, \nabla^2 \psi)$ is Jacobian and $\beta = \frac{\partial f}{\partial y}$ is the beta parameter. If Eq. (2.34) is differentiated by $\frac{\partial}{\partial x}$ and Eq. (2.33) by $\frac{\partial}{\partial y}$ and takes the sum between these two equations, the following equation is obtained after scaling,

$$\nabla^2 gz = \nabla \cdot f \nabla \psi + 2J \frac{\partial \psi}{\partial x}, \frac{\partial \psi}{\partial y} \tag{2.40}$$

This is the non-linear balance equation. It is used in initialization for adjusting the relationship between wind field and geopotential height to satisfy the assumption in SILEPE model (Krishnamurtri, 1986).

2.4 Ensemble Forecasting

In numerical weather prediction, any small error in the initial condition many leads to amplification of the error in forecasting. All scales of motion are affected by the initial error leading to total loss of predictive information. Ensemble forecast technique is introduced in such a way that the initial perturbed state of the ensemble members will be defined. Instead of running just a single forecast, the model is run a number of times from slightly different starting initial conditions. The initial condition differences between ensemble members are very small so that it would be impossible to say which members fitted the observations better. Figure 2.2 shows divergence of forecast in an ensemble with slightly difference initial condition. The heavy line represents the ordinary deterministic forecast starting from an analysis (the cross). At the intermediate forecast projection, the members are still close to one another. At the final forecast projection, clusters (or regimes) may be formed and one of them probably contains the true state of the atmosphere (Wilks, 1995).

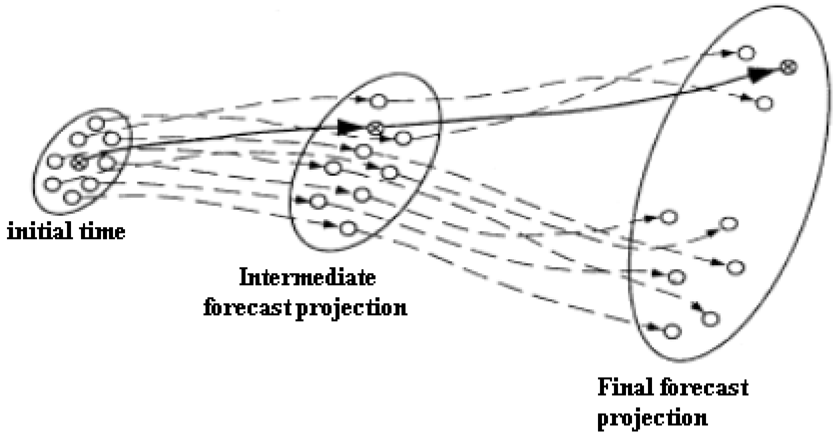

Figure 2.2 Schematic diagram illustrating divergence of forecasts in an ensemble with slightly different initial conditions (Wilks, 1995).

2.4.1 Singular Vector Method

Singular vectors (SVs) method is explained according to Kalnay (2003). The initial conditions for the ensemble prediction system (EPS) are created to represent the uncertainties in the operational analysis. They are obtained by adding to the operation analysis perturbations which produce the fastest growth in the first stage of the forecast period, defined using the singular vector technique.

The objective is to find those perturbations to a given initial state which grow most rapidly. The problem can be formalized mathematically and solutions found if certain assumptions are created. The main assumption is the perturbations grow linearly in time. The singular vectors are the perturbations with the greatest linear growth over the specified time interval for given norms and target areas. Different norms may lead to different sets of singular vectors.

The calculation of the singular vectors becomes feasible if the evolution of the perturbations is linear. In the atmosphere this assumption is generally valid for small perturbations for one or two days ahead. To determine the growth of a perturbation over time a forecast model is needed, but this model must be specifically designed for the linear evolution of small perturbations. Technically this model is the tangent-linear version of the full (non -linear) operational forecast model. A second model, the adjoint of the tangent linear model is also needed in the singular vector calculation.

2.4.1.1 Tangent Linear Model (TLM)

Consider a nonlinear model which has been discretized in space leading to n independent variables, the model can be written as a set of n nonlinear coupled ordinary differential equations

$$\frac{d\mathbf{x}}{dt} = \mathbf{F}(x) \tag{2.41}$$

where $\mathbf{x} = \begin{bmatrix} x_1 \\ \vdots \\ x_n \end{bmatrix}$, $\mathbf{F} = \begin{bmatrix} F_1 \\ \vdots \\ F_n \end{bmatrix}$.

This is the model in differential form. Once a time-difference scheme is applied, it becomes a set of non-linear coupled difference equations.

A numerical solution starting from an initial time t_0 can be obtained by integrating the model numerically between t_0 and a final time t. This gives a nonlinear model solution that depends only on initial conditions

$$\mathbf{x}(t) = \mathbf{M}[x(t_0)] \tag{2.42}$$

where $\mathbf{x}(t_0)$ is the state at the initial time and $\mathbf{x}(t)$ is the state at time t.

$\mathbf{M}$ is the time integration of the numerical scheme from the initial condition to time t.

The Tangent Linear Model (TLM), is defined such that

$$\mathbf{L} = \frac{\partial \mathbf{M}}{\partial x} \tag{2.43}$$

where $\mathbf{L}$ is the propagator of the tangent linear model. It propagates an initial perturbation at time t_0 to final perturbation at time t.
So the TLM evolves a perturbation in time

$$\mathbf{y}(t) = \mathbf{L}(t_0, t)\mathbf{y}(t_0) \tag{2.44}$$

where $\mathbf{y}(t_0)$ is the perturbation state at the initial time $t = 0$ and $\mathbf{y}(t)$ is the perturbation state at time t.

2.4.1.2 The Adjoint Model

The adjoint tangent linear model is the transpose of the tangent linear model. It is defined with respect to the inner product of two arbitrary vectors

$$\langle \mathbf{L}u, v \rangle = \langle u, \mathbf{L}^T v \rangle \tag{2.45}$$

where u and v are arbitrary vectors and $\mathbf{L}^{\mathbf{T}}$ is the adjoint TLM.

2.4.1.3 Singular Vectors

Most of the content in this section follows Kalnay (2003). For an interval (t_0, t_1), the TLM is a matrix that when applied to a small initial perturbation $\mathbf{y}(t_0)$ produces the final perturbation $\mathbf{y}(t_1)$

$$\mathbf{y}(t_1) = \mathbf{L}(t_0, t_1)\mathbf{y}(t_0). \tag{2.46}$$

Singular value decomposition theory indicates that for any matrix $\mathbf{L}$ there exist two orthogonal matrices $\mathbf{U,V}$ such that

$$\mathbf{U}^T\mathbf{LV} = \mathbf{S} \tag{2.47}$$

where $\mathbf{S} = \begin{bmatrix} \sigma_1 & 0 & \dots & 0 \\ 0 & \sigma_2 & \dots & 0 \\ \vdots & \vdots & & \vdots \\ 0 & 0 & \dots & \sigma_n \end{bmatrix}$ and $\mathbf{UU}^T = \mathbf{I}$, $\mathbf{VV}^T = \mathbf{I}$

$\mathbf{S}$ is a diagonal matrix whose elements (σ_i) are the singular values of $\mathbf{L}$.

$\mathbf{I}$ is the identity matrix.

Left multiply Eq. 2.47 by $\mathbf{U}$, to get

$$\mathbf{LV} = \mathbf{US} \tag{2.48}$$

$$\mathbf{Lv}_i = \sigma_i \mathbf{u}_i \tag{2.49}$$

where $\mathbf{v}_i$ are the columns of $\mathbf{V}$ and $\mathbf{u}_i$ are the columns of $\mathbf{U}$.

Eq. 2.49 defines $(\mathbf{v_1,\ldots,v_n})$ as the right singular vectors of $\mathbf{L}$, hereafter referred to as initial singular vectors, since they are valid at the beginning of optimization interval over which $\mathbf{L}$ is defined.

$$\mathbf{L}^T\mathbf{U} = \mathbf{VS} \tag{2.50}$$

$$\mathbf{L}^T\mathbf{u}_i = \sigma_i \mathbf{v}_i \tag{2.51}$$

The velocity $(\mathbf{u_1,\ldots,u_n})$ are the left singular vectors of $\mathbf{L}$ and will be referred to as final singular vectors, since they correspond to the end of the interval of optimization.

Consequently, the equation is derived as

$$\mathbf{LL^T u}_i = \sigma_i(\sigma_i \mathbf{u}_i) = \sigma_i^2 \mathbf{u}_i \tag{2.52}$$

$$\mathbf{L}^T\mathbf{L}\mathbf{v}_i = \sigma_i^2 \mathbf{v}_i \tag{2.53}$$

Therefore, the initial singular vectors can be obtained as the eigenvectors of $\mathbf{L}^T\mathbf{L}$, the final singular vectors can be obtained as the eigenvectors of $\mathbf{LL}^T$ and a normal matrix whose eigenvalues are the squares of the singular values.

Considering a basic trajectory

$$\mathbf{y}(t_0) = \sum_{i=1}^{n} < \mathbf{y}(t_0), \mathbf{v}_i > \mathbf{v}_i \tag{2.54}$$

$$\mathbf{y}(t_1) = \sum_{i=1}^{n} < \mathbf{y}(t_1), \mathbf{u}_i > \mathbf{u}_i \tag{2.55}$$

$$\mathbf{y}(t_1) = \mathbf{L}(t_0, t_1) \sum_{i=1}^{n} < \mathbf{y}(t_0), \mathbf{v}_i > \mathbf{v}_i \tag{2.56}$$

$$< \mathbf{y}(t_1), \sigma_i \mathbf{u}_i > = < \mathbf{y}(t_0), \mathbf{v}_i >. \tag{2.57}$$

This indicates that by applying the tangent linear model $\mathbf{L}$ each initial vector $\mathbf{v}_i$ component will be stretched by an amount equal to the singular value σ_i and the direction will be rotated to that of the evolved vector $\mathbf{u}_i$ as shown in Figure 2.3.

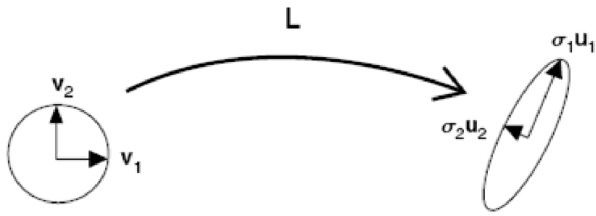

Figure 2.3 Schematic of the application of the tangent linear model to a sphere of perturbations of size 1 for a given interval (t_0, t_1) (Kalnay, 2003).

From Eq. 2.55 and taken $\mathbf{L}^T$ into the equation,

$$\mathbf{L}^T \mathbf{y}(t_1) = \sum_{i=1}^{n} < \mathbf{y}(t_0), \mathbf{v}_i > \sigma_i \left(\mathbf{L}^T \mathbf{u}_i \right) \tag{2.58}$$

From Eq. 2.51; $\mathbf{L}^T \mathbf{u}_i = \sigma_i \mathbf{v}_i$

$$\mathbf{L}^T \mathbf{y}(t_1) = \sum_{i=1}^{n} < \mathbf{y}(t_0), \mathbf{v}_i > \sigma_i^2 \mathbf{v}_i \tag{2.59}$$

By property of inner product

$$< \mathbf{y}(t_1), \sigma_i \mathbf{u}_i > = \sigma_i^2 < \mathbf{y}(t_0), \mathbf{v}_i > \tag{2.60}$$

This indicates that by applying $\mathbf{L}^T \mathbf{L}$ is running tangent linear model forward in time, and then the adjoint backward in time, the first initial singular vectors $\mathbf{v}_i$ will grow by factor σ_i^2 as shown in Figure 2.4.

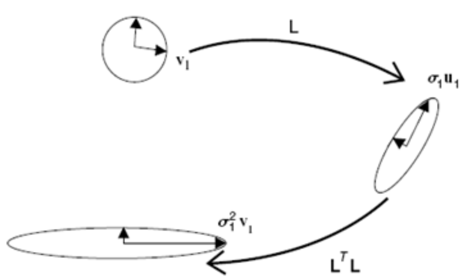

Figure 2.4 Schematic of the application of the tangent linear model forward in time followed by the adjoint of the tangent linear model to a sphere of perturbations of size 1 at the initial time. (Kalnay, 2003).

2.4.2 Monte Carlo Method

The Monte Carlo method for ensemble forecasting repeats random perturbations of all data simultaneously. The resulting forecasts will differ almost throughout the forecast domain. The idea of this method is repeating the experiment many times with different sets of number, the forecast errors that are due to the uncertainty of the observations and analyses can be obtained. (Krishnamurti et.al, 2000)

2.4.3 Breeding Method

Breeding (Toth and Kalnay, 1993, 1997, Kalnay et al., 2002) is developed as a method to generate initial perturbations for ensemble forecasting at the National Centers for Environmental Prediction (NCEP). The method involves simply running the nonlinear model used for the control a second time, periodically subtracting the control from the perturbed solution, and rescaling the difference so that it has the same size as the original perturbation. The rescaled difference (a bred vector) is added to the control run and the process repeated. In the context of data assimilation, the rescaled difference is added to the analysis (Figure 2.5). Bred vectors are a nonlinear generalization of leading Lyapunov vectors. Their growth rate is a measure of the local instability of the flow.

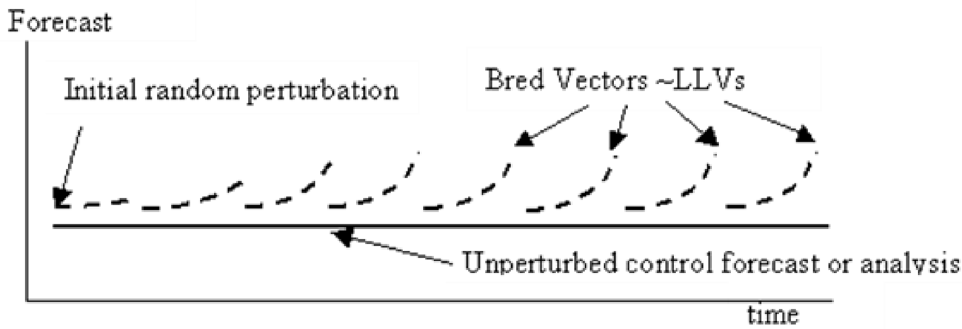

Figure 2.5 Schematic of the method to generate bred vectors (Toth and Kalnay, 1993).

The breeding method procedure consists of the following steps:

1. Add a small arbitrary perturbation to the atmospheric analysis.
2. Integrate the model for a short period (e.g. 6 hours) from both the unperturbed (control) and the perturbed initial conditions.
3. Subtract the control forecast from the perturbed forecast.

4. Scale down the difference field so that it has the same size as the initial perturbation.

Figure 2.6 shows that a small arbitrary perturbation is introduced on the control analysis initially. After 6-hr integration, the difference between the control and perturbed forecasts is scaled back to the size of the initial perturbation and this difference field is then added into the new analysis. After 3–4 days of cycling, the perturbation is dominated by growing modes due to the 'natural selection' of fast-growing perturbations (Toth and Kalnay, 1993).

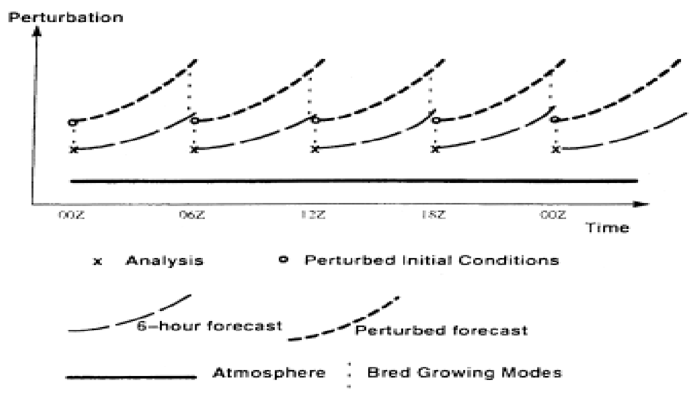

Figure 2.6 Schematic illustration of the 6-hr breeding cycle (Toth and Kalnay, 1993).

2.5 Active and Break Summer Monsoon

Active and break events are identified from the average rainfall data over a critical area, called the core monsoon zone within which the monsoon trough/ Continental Tropical Convergence Zone (CTCZ) normally fluctuates in the peak monsoon months of July and August (Takahashi and Yasunari, 2006). Thai Meteorological Department (2012) defines the break in summer monsoon as the period in rainy season in which the rainfall is less than 0.1 mm continuously for 15 days. Sukuwat (2012) states that the break in summer monsoon can be identified from the Inter Tropical Convergent Zone moves to southern China, the high pressure ride over Australia extends to the west of Indochina and significant decrease in rainfall over Southeast Area.