NOTES AND CORRESPONDENCE

Semigeostrophic Theory

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ABSTRACT

The geopotential tendency form of semigeostrophic theory is derived and compared with the potential vorticity form. The tendency form is compact and particularly convenient for non-Boussinesq, nonuniform potential vorticity flows.

1. Introduction

By combining the geostrophic momentum approximation (Eliassen, 1948) and the geostrophic coordinate transformation, Hoskins (1975) and Hoskins and Draghici (1977) have derived a filtered, three dimensional system of equations (semigeostrophic equations) which are nearly as simple as the quasi-geostrophic equations but which apply to more general physical situations such as fronts, jets, occluding baroclinic waves, etc. With a specified, height independent deformation field, two dimensional versions of the semigeostrophic equations (Hoskins, 1971; Hoskins and Bretherton, 1972) give fairly realistic simulations of both surface fronts (uniform potential vorticity model) and upper-level fronts (discontinuous potential vorticity model). In the three dimensional case uniform potential vorticity jet flows develop unstable baroclinic waves which evolve into the nonlinear regime, producing fronts and an occluding warm sector (Hoskins, 1976; Hoskins and West, 1979; Hoskins and Heckley, 1980). Three dimensional nonuniform potential vorticity jet flows produce similar evolution except that upper tropospheric frontogenesis is more realistic (Heckley and Hoskins, 1982).

In this note we derive the "geopotential tendency" form of semigeostrophic theory. This form is convenient because the ageostrophic circulation remains implicit and only one elliptic equation needs to be solved each time step. From the computational point of view integrations of the nonuniform potential vorticity case become almost as easy as integrations of the uniform potential vorticity case.

2. Review of semigeostrophic theory

a. Geostrophic momentum approximation

Our starting point is the $f$-plane system of equations with the geostrophic momentum approximation [see Eqs. (1)-(5) of Table 1]. Here we use the pseudo-height $z = [1 - (p/p_0)]^2 e p_0 / g$ as the vertical coordinate. The pseudo-density $\rho = \rho_0 [1 - gz/c_0]\theta_0^{(1-\nu)e}$ is then a known function of $z$. In (1)-(5), $(u_g, v_g) = f^{-1}(-\partial \phi/\partial y, \partial \phi/\partial x)$ are the geostrophic components of the wind, $(u_a, v_a)$ the ageostrophic components, and $(u, v) = (u_g + u_a, v_g + v_a)$ the components of the total wind.

Although the derivation is lengthy, it can be shown that the above system has the three-dimensional vorticity equation

$$\frac{D \zeta}{Dt} = \left( \frac{\zeta \cdot \nabla }{\rho} \right) u + \frac{\zeta}{\rho \zeta} \frac{\partial \rho}{\partial z} - \frac{g}{\theta_0} \mathbf{k} \times \nabla \theta,$$

(6)

where $u = (u, v, w)$, $D/Dt = \partial / \partial t + u \partial / \partial x + v \partial / \partial y + w \partial / \partial z$, and

$$\zeta = \left( \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) + \left( \frac{\partial u}{\partial y} - \frac{\partial u}{\partial x} \right) + \frac{1}{f} \frac{\partial \phi}{\partial y}$$

$$+ \frac{1}{f} \frac{\partial \phi}{\partial x}.$$

Combining (6) with the continuity equation (4) and the thermodynamic equation (5) we obtain the potential vorticity equation

$$\frac{D q}{Dt} = 0,$$

(8)

where

$$q = \frac{g}{\rho_0 f} \zeta \cdot \nabla \theta.$$

(9)

b. Semigeostrophic equations

Introducing the geostrophic coordinates

$$(X, Y, Z, T) = \left( x + \frac{u_g}{f}, y - \frac{u_g}{f}, z, t \right)$$

(10)

and the potential function

$$\Phi = \phi + \frac{1}{2} (u_g^2 + v_g^2),$$

(11)

Hoskins (1975) has shown that the geostrophic and hydrostatic relations take the form...
Table 1. Semigeostrophic theory: Equations.

<table>
<thead>
<tr>
<th>Geostrophic momentum approximation</th>
<th>Semigeostrophic equations</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - f u = 0$</td>
<td>$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} - f u^* = 0$</td>
</tr>
<tr>
<td>$\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + f v = 0$</td>
<td>$\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + f v^* = 0$</td>
</tr>
<tr>
<td>$\frac{\partial \phi}{\partial \tau} = \frac{g}{\theta_0} \theta$</td>
<td>$\frac{\partial \phi}{\partial \tau} = \frac{g}{\theta_0} \theta^*$</td>
</tr>
<tr>
<td>$\frac{\partial \theta}{\partial \tau} + u \frac{\partial \theta}{\partial x} + v \frac{\partial \theta}{\partial y} + w \frac{\partial \theta}{\partial z} = 0$</td>
<td>$\frac{\partial \theta^<em>}{\partial \tau} + u \frac{\partial \theta^</em>}{\partial x} + v \frac{\partial \theta^<em>}{\partial y} + f \frac{\partial \theta^</em>}{\partial \theta} = 0$</td>
</tr>
</tbody>
</table>

while the potential vorticity becomes

$$q = \frac{g}{\rho_0} \frac{\zeta}{f} \frac{\partial \theta}{\partial Z},$$

where $\zeta$ is the vertical component of absolute vorticity. Since $D/DT = \partial/\partial T + u \partial/\partial X + v \partial/\partial Y + w \partial/\partial Z$, the horizontal advection is now geostrophic. Then, introducing the transformed geostrophic components

$$(u^*, v^*, w^*) = \left( u + \frac{w}{f} \frac{\partial v}{\partial z}, v - \frac{w}{f} \frac{\partial u}{\partial z}, \frac{f}{g} w \right),$$

Hoskins and Dranghici (1977) showed that (1)–(5) can be written as (15)–(19) of Table 1.

Although the system (15)–(19) can be regarded as closed, it is not convenient for calculation. The geostrophic components $(u^*, v^*, w^*)$ are implicit and must be determined in such a way that the predictions of $(u_0, v_0)$ from (15)–(16) and $\theta$ from (19) are consistent with a continued state of geostrophic and hydrostatic balance. In Sections 3 and 4 we discuss two alternate forms of the semigeostrophic equations.

3. Potential vorticity form

From (15) and (16) we can obtain

$$D_x \left(\frac{g}{\theta_0} \nabla \theta \right) + \nabla (p w^*) = \frac{1}{2} Q,$$

where $D_x = \partial/\partial T + u_0 (\partial/\partial X) + v_0 (\partial/\partial Y)$ and $Q = -2g/\theta_0 \partial \phi_0/\partial X \cdot \nabla \theta$, $\partial \phi_0/\partial Y \cdot \nabla \theta$, with $\nabla$ denoting the horizontal del operator in geostrophic space. From (19) we can also obtain

$$\begin{align*}
\frac{\partial \theta}{\partial \tau} + u \frac{\partial \theta}{\partial x} + v \frac{\partial \theta}{\partial y} + w \frac{\partial \theta}{\partial z} = 0
\end{align*}$$

since the thermal wind equation [derived from (12)] can be written

$$f k \times \frac{\partial \theta}{\partial Z} + \frac{g}{\theta_0} \nabla \theta = 0,$$

(20) and (21) can be added to yield

$$\nabla (p w^*) - f^2 \frac{\partial \theta^*}{\partial \theta} = Q.$$  

Taking the horizontal divergence of (23) and combining the result with the continuity equation (18) we obtain (24), the $w^*$ equation given at the top of the left column in Table 2.

Using (12)–(14) the conservation of potential vorticity can be written as (25). Finally, the relation between $q$ and $\Phi$, and the thermodynamic equation applied at the boundary can be written as (26). Equations (24)–(26) constitute a closed set in the unknowns $\Phi$, $q$, $w^*$. This is the set discussed by Hoskins (1982, p. 145) and used by Heckley and Hoskins (1982). Numerical integration proceeds in the order given in Table 2. Note that two elliptic equations (one for $w^*$, one for $\Phi$) must be solved each time step.

There is a special case which results in considerable simplification of (24)–(26). This is the case of uniform potential vorticity flow. In this case $w^*$ need not be computed from (24), and $q$ in (26) is a constant. Thus, the whole dynamics reduces to (26). The Boussinesq approximation ($\rho = $ constant) leads to even further simplification. A great deal can be learned about fronts and baroclinic waves using this Boussinesq, uniform potential vorticity model (e.g., Hoskins, 1976; Hoskins and West, 1979).

4. Geopotential tendency form

We now derive the form of semigeostrophic theory which involves $\Phi_T$. Since $\Phi_T = \phi_1$ we call this the geopotential tendency form.

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1 Hoskins and Dranghici (1977) use the Boussinesq form of (18), but their analysis easily generalizes to the non-Boussinesq case given in Table 1.
TABLE 2. Potential vorticity and geopotential tendency forms of semigeostrophic theory.

<table>
<thead>
<tr>
<th>Potential vorticity form</th>
<th>Geopotential tendency form</th>
</tr>
</thead>
<tbody>
<tr>
<td>[ \nabla \times (\rho q v^w) + f^2 \frac{\partial}{\partial Z} \left( \frac{\partial q}{\partial Z} \right) = \nabla \cdot Q ]</td>
<td>[ \rho q = \frac{f^3}{f^2 - (\Phi_T + \Phi_R)} ]</td>
</tr>
<tr>
<td>B.C. ( w^* = 0 ) at ( Z = 0, Z_T )</td>
<td>[ \frac{f^3}{f^2 - (\Phi_T + \Phi_R)} ]</td>
</tr>
<tr>
<td>[ \left( \frac{\partial}{\partial T} - \frac{1}{f} \frac{\partial}{\partial Y} \right) \left( \frac{\partial}{\partial Z} \left( \frac{\partial}{\partial Z} \right) \right) + \frac{\partial}{\partial Z} \left( \frac{f^2}{q} D \frac{\partial}{\partial Z} \right) = 0 ]</td>
<td>[ \text{B.C.} \left( \frac{\partial}{\partial T} - \frac{1}{f} \frac{\partial}{\partial Y} \right) \left( \frac{\partial}{\partial Z} \left( \frac{\partial}{\partial Z} \right) \right) ]</td>
</tr>
<tr>
<td>at ( Z = 0, Z_T )</td>
<td>( \frac{f^3}{f^2 - (\Phi_T + \Phi_R)} )</td>
</tr>
</tbody>
</table>

Adding \( \partial (\rho f^2 (16)/\partial X, \partial (-\rho f^2 (15)/\partial Y, \partial f^2 (19)/\partial Z) \), then using (12) and (18) we obtain

\[ \frac{\partial}{\partial X} \left[ \rho D_e \left( \frac{\partial \Phi}{\partial X} \right) \right] + \frac{\partial}{\partial Y} \left[ \rho D_e \left( \frac{\partial \Phi}{\partial Y} \right) \right] + \frac{\partial}{\partial Z} \left[ \frac{f^2}{q} D_e \left( \frac{\partial \Phi}{\partial Z} \right) \right] = 0. \] (27)

Defining

\[ P = \left( \rho \frac{\partial \Phi}{\partial X}, \nabla \Phi, \rho \frac{\partial \Phi}{\partial Y}, \nabla \Phi, \frac{f^2}{q} \frac{\partial \Phi}{\partial Z}, \nabla \Phi \right), \] (28)

(27) can be written as Eq. (30) of Table 2. Equation (26) can now be rearranged to the form given by (29). Equations (29) and (30) constitute a closed set in \( \Phi_T \) and \( q \). Only one elliptic equation needs to be solved each time step. The ageostrophic circulation remains entirely implicit. If \( w^* \) is desired, it can be easily computed from (19) using the known \( \Phi_T \). In the special case of uniform potential vorticity flow it can be shown that \( \nabla \cdot P = 0 \), so that (26) and (30) become essentially equivalent.

Equation (30) is somewhat analogous to an equation derived by Eliassen (1974) in a \( \theta \)-coordinate type model with the geostrophic momentum approximation. The relation between (30) and the potential vorticity equation (25) has been discussed in the context of Eliassen's balanced vortex model (Schubert and Hack, 1983; Schubert et al., 1983) by Thorpe (1985).

5. Concluding remarks

The main result obtained here is the right-hand side of Table 2. This geopotential tendency form of semigeostrophic theory is compact and allows numerical integration of non-Boussinesq, nonlinear potential vorticity flows to be implemented almost as simply as the Boussinesq, uniform potential vorticity case. The word "almost" must be used because for the Boussinesq, uniform potential vorticity case the interior part of (26) can be solved analytically and numerical integration need be implemented only on the boundary condition part of (26), as described by Hoskins (1976); in contrast, the non-Boussinesq, nonlinear potential vorticity case requires that we numerically solve the interior part of (30). Although repeatedly solving three dimensional elliptic problems can be time consuming, fast multigrid solvers are becoming available. Finally we note that the distinction between the potential vorticity form and the geopotential tendency form disappears when the additional quasi-geostrophic approximations are introduced into (1)–(5).

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