Surface Kinetic Energy Transfer in SQG Flows

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(Received 9 July 2007)

The relevance of surface quasi-geostrophic dynamics (SQG) to the upper ocean and the atmospheric tropopause has been recently demonstrated in a wide range of conditions. Within this context, the properties of SQG in terms of kinetic energy (KE) transfers are revisited and further explored. Two well-known properties of SQG in spectral space are (i) the identity between surface velocity and density spectra (when appropriately scaled) and (ii) the existence of a forward cascade for surface density variance. Here we show numerically and analytically that (i) and (ii) do not imply a forward cascade of surface KE (through advection). On the contrary, advection by the geostrophic flow primarily induces an inverse cascade of surface KE on a large range of scales. This spectral flux is locally compensated for by a KE source, linked to surface frontogenesis, \textit{that is interpreted as a conversion of potential to kinetic energy}. The subsequent spectral budget resembles those exhibited by more complex systems (primitive equation or Boussinesq models) and observations, which strengthens the relevance of SQG for the description of ocean/atmosphere dynamics near vertical boundaries. The main weakness of SQG however is in the small-scale range (scales smaller than 20-30km in the ocean) where it poorly represents the forward KE cascade observed in non-QG numerical simulations.

1. Introduction

A fundamental question of ocean and atmosphere dynamics is how their equilibrium energy spectrum is established and what the underlying spectral energy transfers are. \textbf{A variety of dynamical regimes must be distinguished}. In particular regions close to boundaries such as the ocean surface or the atmospheric tropopause behave differently than the interior. For boundaries Blumen (1978) developed a surface quasi-geostrophic (SQG) theory that serves as a counterpart to the model of 3-D geostrophic turbulence (Charney 1971). This latter is based on the quasi-geostrophic (QG) nature of large-scale interior flow and is not influenced by boundary contrasts. The SQG dynamics on the other hand is entirely driven by the density (or potential temperature in the atmosphere) anomaly evolution at the boundary. As such frontogenesis (in its QG limit) is the key process in SQG systems. SQG theory has been recently used to describe the 3-D dynamics of the upper troposphere (Juckes 1994; Hakim \textit{et al.} 2002; Tulloch & Smith 2006) and
the upper oceanic layers (Lapeyre & Klein 2006a; LaCasce & Mahadevan 2006; Isern-Fontanet et al. 2006). In order to better understand the range of applicability of the SQG theory, Lapeyre & Klein (2006a) have further revisited theoretically and numerically the question of the coupling of the boundary with the interior. Noting the dynamical analogy (suggested by Bretherton (1966)) of the surface density as a boundary potential vorticity (PV) delta-function and using the invertibility principle of PV (Hoskins et al. 1985), Lapeyre & Klein (2006a) demonstrate the relevance of the SQG dynamics in the upper oceanic layers for a spectral range extending from the smallest scales ($O(10\,\text{km})$) to mesoscales (as large as $O(500\,\text{km})$), the larger scales being predominantly influenced by the interior PV. Such scale partition also emerges from the studies of Juckes (1994) and Tulloch & Smith (2006) for atmospheric flows.

The spectral energy transfers at the boundaries have still to be understood. Using altimetry data Scott & Wang (2005) computed KE fluxes due to nonlinear horizontal advection. They found an inverse cascade of surface kinetic energy (KE) from scales close to the Rossby deformation radius ($\approx 100 - 150\,\text{km}$ at mid-latitudes) towards larger scales. Their interpretation is that this inverse cascade mainly reflect the first baroclinic mode. In support of this hypothesis Scott & Arbic (2007) present two-layer QG turbulent solutions for a baroclinically unstable ocean that undergoes an inverse cascade of the baroclinic (and upper layer) KE. Using primitive equation (PE) simulations Capet et al. (2007) and Klein et al. (2007), confirm Scott & Wang (2005) results about the existence of a significant inverse cascade of surface KE. Their inverse cascade, again estimated from the nonlinear horizontal advection terms, extends even further down to scales around 30km. It is not clear that this cascade can be understood in terms of QG turbulence that assumes uniform surface density. Indeed, in agreement with Lapeyre & Klein (2006a), density contrasts at the ocean surface play a leading dynamical role for scales up to at least 300km (Klein et al. 2007). In this context we revisit SQG energy transfers and show that they account for an inverse surface KE cascade at least within the mesoscales to small scales range.

Blumen (1978) and later Held et al. (1995) noted two invariants in SQG dynamics, the surface density variance (also equal to the surface KE when appropriately dimensionalized) and the depth-integrated total energy (KE plus potential energy). By analogy with 2-D turbulence the system undergoes a direct cascade of surface density variance at high wavenumbers and an inverse cascade of the total energy at low wavenumbers. Neither Blumen (1978) nor Held et al. (1995) discussed the specific properties of the surface KE cascade in SQG theory. The focus of this paper is to characterize these properties. In sections 2 and 3 we show (first analytically and then numerically) that despite the surface density and velocity spectra being identical in SQG, the details of their spectral budget differ: horizontal advection induces an upscale flux of surface KE over a large spectral range extending to small scales whereas it fluxes density variance downscale. The difference is compensated for by an extra term in the surface KE spectral budget that we connect to frontogenesis in section 4. The conclusion (section 5) examines the relevance and limitations of the SQG framework to understand energy spectral transfers in more complex systems.

2. Surface KE and density variance budgets

2.1. The SQG equations

The SQG model (see Blumen (1978) and Held et al. (1995)) is based on a small Rossby number approximation to the primitive equations (that describe an hydrostatic Boussi-
nesq fluid system). All variables are nondimensionalized as in Pedlosky (1987) using $U$ and $L$ respectively the velocity and length scales, $f$ the Coriolis parameter, $N$ the Brunt-Väisälä frequency and $H$ the depth scale. The Rossby number (defined as $\epsilon = U/fL$) is assumed small ($\epsilon << 1$) and the Burger number (defined as $B = NH/fL$) is of order one ($B = 1$). For any variable $X$, its perturbation expansion in terms of the Rossby number $\epsilon$ writes:

\[
X = X^0 + \epsilon X^1 + \mathcal{O}(\epsilon^2).
\]

Retaining the $\mathcal{O}(1)$-terms in the momentum and hydrostatic balances leads to the resulting geostrophic and hydrostatic relations

\[
(u^0, v^0, \rho^0) = (-\Phi^0_y, \Phi^0_x, -\Phi^0_z)
\]

where $u$ and $v$ are the horizontal velocity components respectively along the zonal ($x$) and meridional ($y$) coordinates. $\rho$ is the density and $z$ the vertical coordinate (with $z < 0$ for an oceanic setup). The streamfunction $\Phi^0$ is related to pressure. SQG theory furthermore assumes that potential vorticity (PV) is uniform in the interior of the fluid and the flow decays away from the surface (i.e., $\Phi^0 \to 0$ as $z \to -\infty$). The resulting PV equation in nondimensional form is

\[
\nabla_H^2 \Phi^0 + \partial_{zz} \Phi^0 = 0,
\]

with the boundary condition

\[
-\partial_z \Phi^0 |_{z=0} = \rho_s.
\]

The subscript $s$ refers to surface variables and $\nabla_H$ the horizontal gradient operator. Solving these two equations in the spectral Fourier space leads to

\[
\hat{\Phi}^0(k, l, z) = -\frac{\hat{\rho}_s(k, l)}{K} \exp(Kz),
\]

where $\hat{\cdot}$ is the horizontal spectral transform. $k, l$ are the horizontal wavenumbers along the $x$ and $y$ directions and $K = (k^2 + l^2)^{1/2}$. The dynamics at zero order and at any depth is thus entirely determined by the surface density, $\rho_s$. Time evolution of the flow requires one to consider the next leading order approximation (in $\epsilon$) to the primitive equations and in particular that for surface density:

\[
\partial_t \rho_s + u^0_s \cdot \nabla_H \rho_s = 0,
\]

with $t$ the time. (2.1), (2.4) and (2.5) form a closed system, i.e., the basic SQG equation set.

Using the density equation (and retaining only the $\mathcal{O}(\epsilon)$-terms), we have at any depth

\[
w^1 = \partial_t \rho^0 + u^0 \cdot \nabla_H \rho^0.
\]

Thus, from (2.4), (2.5) and (2.6), the vertical velocity in spectral space can be computed diagnostically

\[
\hat{w}^1 = -\nabla_H \cdot (u^0_s \rho_s) \exp(Kz) + \nabla_H \cdot (u^0 \rho^0).
\]

where $\hat{\rho}^0(k, l, z)$ is given by $\hat{\rho}^0(k, l, z) = \hat{\rho}_s(k, l) \exp(Kz)$ (using (2.1) and (2.4)). Similarly, taking the $z$-derivative of (2.6) and using (2.1) and (2.4) provides a diagnostic expression for $w^1_z$ as a function of surface density

\[
\hat{w}^1_z = -K \nabla_H \cdot (u^0_s \rho_s) \exp(Kz) + \nabla_H \cdot (u^0 \rho^0_z).
\]

Because of the hypothesis of uniform interior PV horizontal and vertical structures of the flow are related to each other in this balanced model. This explains why the horizontal as well as vertical velocity fields can be retrieved at all depths from surface density only (Hakim et al. 2002; Lapeyre & Klein 2006a). Another important property within the
context of this study is that surface velocity and density spectra are identical. Indeed from (2.1) and (2.4), we have

$$|\hat{\rho}_s|^2(k, l) = |\hat{u}^0_s|^2(k, l).$$

(2.9)

Finally it should be noted that a difference between SQG and 2-D flows emerges from the form of the preceding equations. In SQG flows the conserved scalar is the surface density (2.5) instead of the relative vorticity, $Z = v_x - u_y$, for 2-D flows. The streamfunction at the surface is given by $\Phi^0_s(k, l) = \frac{\tilde{Z}(k, l)}{K}$ (see (2.4)) instead of $\Phi^0(k, l) = -\frac{\tilde{Z}(k, l)}{K}$ for 2-D flows. This implies that near the surface large-scale strain in SQG plays a relatively lesser role in the advection of small-scale features, resulting in a cascade of density variance to small scales that is more local in wavenumber (Pierrehumbert et al. 1994).

2.2. KE and density variance equations

Time evolution of the horizontal velocity, within the QG framework, is obtained by retaining only the $O(\epsilon)$-terms in the momentum equations (Pedlosky 1987). At the surface the resulting equation is

$$\partial_t u^0_s = -\nabla^s u^0_s - k \wedge u^{ag}_s$$

(2.10)

where $k$ is the unit vertical vector and $u^{ag}_s = u^1_s + k \wedge \Phi^1_s$ the horizontal velocity ageostrophic component. We neglect dissipation in (2.10). By spectrally manipulating (2.5) and (2.10), we get the surface kinetic energy and density variance budgets in spectral space:

$$\partial_t |\hat{\rho}_s|^2/2 = -Re[\hat{\rho}_s \cdot (\hat{u}^0_s \cdot \nabla \hat{\rho}_s)]$$

(2.11)

$$\partial_t |\hat{u}^0_s|^2/2 = -Re[\hat{u}^0_s \cdot (\hat{u}^0_s \cdot \nabla \hat{u}^0_s)] - Re[\hat{w}^1_z \Phi^0_s],$$

(2.12)

where * is the conjugate. In (2.12), we have utilized the identities

$$\hat{u}^0_s \cdot k = \hat{u}^{ag}_s \cdot \nabla \Phi^0_s = -\nabla \cdot \hat{u}^{ag}_s \Phi^0_s = w^1_z \Phi^0_s,$$

noting $\nabla \cdot \hat{u}^{ag}_s = w^1_z$.

The nonlinear transfers due to the horizontal advection are captured by the first RHS term in (2.11) and (2.12). Integration of these equations with respect to $K$ and averaging isotropically (noted as $< . . >$) leads one to introduce the density variance and kinetic energy spectral fluxes related to those terms, i.e. respectively

$$\Pi_\rho = -\int_K < \Re[\hat{\rho}^*_s \cdot \hat{u}^0_s \cdot \nabla \hat{\rho}_s] > dK$$

(2.13)

and

$$\Pi_u = -\int_K < \Re[\hat{u}^0_s \cdot \hat{u}^0_s \cdot \nabla \hat{u}^0_s] > dK.$$  

(2.14)

We also define

$$\Pi_a = -\int_K < \Re[\hat{w}^1_z \Phi^0_s] > dK,$$

(2.15)

related to the last term in the RHS of (2.12) and that involves ageostrophic variables. Given (2.9) the RHS of (2.11) and (2.12) are equal which yields

$$\Pi_\rho = \Pi_u + \Pi_a.$$  

(2.16)

The relation (2.16) states that the surface density variance and KE fluxes due to geostrophic advection (respectively $\Pi_\rho$ and $\Pi_u$) differ by $\Pi_a$ that involves the ageostrophic flow. The
striking consequence of this last term, as shown in the next section, is that the spectral transfer properties of surface KE and of density variance totally differ from each other despite (2.9).

3. Numerical simulations

3.1. Description

We have performed numerical simulations for a SQG turbulent eddy field in free decay. The computation domain is doubly-periodic. The numerics is based on the model of Hua & Haidvogel (1986) modified to solve the SQG equations set. The resolutions we employ are $512^2$ (LR), $1024^2$ (MR) and $2048^2$ (HR). Initial conditions are constructed by specifying a $\hat{\phi}^0$ field at the surface, with random phase angles and amplitude given by $|\hat{\phi}^0|^2(k, l) = \frac{K^5}{(K + ko)^2}$. The wavenumber $k_o = 14$ corresponds to the KE peak. Dissipation involves a biharmonic operator and a coefficient adjusted for each resolution to the minimal value for which no density variance accumulates at small scale (by inspecting the spectra).

The numerical solutions exhibit the well-known features of SQG turbulence. In the density field (see Fig. 1a) small scales are energetic. This is due to frontogenesis that intensifies the density gradients and to destabilization of filaments that leads to numerous eddies having a diameter around a dozen grid points. Consequently, the density spectrum is quite flat, with a slope between $-5/3$ and $-2$ over one and half decade (see Fig. 1b). Note that this description as well as the analyses performed below are for intermediate times, after the spin up was achieved and while the surface flow is composed of numerous eddies and yields active nonlinear interactions. (At a much longer time our free-decay solutions evolve toward a state where the total energy is diminished, surface spectra are steeper than $-2$, and nonlinear interactions are very weak because large isolated eddies dominate the flow.)

3.2. Spectral transfers of surface KE and density variance

The different terms (2.13), (2.14) and (2.15) are computed from 10 snapshots of density (taken over the six nondimensional time units during which an inertial range is most cleanly present; this would correspond to 10 inertial periods for the oceanically relevant
scaling used in (Lapeyre & Klein 2006b). $w_2^1$ (required to compute $\Pi_a$) is diagnosed from (2.8). Results are given for HR in Fig. 2a. The forward cascade of density variance extends roughly from $k = 15$ to the dissipation scale ($k > 500$). This is the classical result anticipated by Blumen (1978) and subsequently verified by numerous studies (see Celani et al. (2004)). The flux is close to constant over this interval. The advective flux of surface KE, on the other hand, is upscale ($\Pi_u < 0$) over a large spectral range. To our knowledge this important aspect of SQG turbulence had been overlooked. A downscale flux is also present but it concerns a narrow range of very small scales (with wavenumbers larger than $k = 300$). From (2.16) the difference between $\Pi_\rho$ and $\Pi_u$ is entirely explained by $\Pi_a$ (whose physical interpretation is discussed in the next section). This has been checked by plotting some values of $\Pi_u + \Pi_a$ in Fig. 2a. Unlike $\Pi_\rho$, $\Pi_u$ and $\Pi_a$ have no plateau.

In Fig. 2b, $\Pi_u$ and $\Pi_\rho$ are plotted for different horizontal resolutions. For $\Pi_\rho$, increasing the resolution simply extends its plateau corresponding to a forward cascade of density variance over a wider spectral range. The same behavior is observed for the the upscale surface KE advective flux, which indicates its robustness. On the other hand, the downscale surface KE flux is not robust with respect to the resolution changes. As resolution increases the region where $\Pi_u$ is positive moves to higher wavenumber and roughly follows the dissipation range. In other words, the downscale advective flux of surface KE would be absent from a solution with infinite resolution.

4. Discussion

Although surface density and velocity spectra are identical over a large spectral range (as confirmed by their tendency terms), their dynamics revealed by their budgets are quite different because of the term $\Pi_a$. The physical meaning of $\Pi_a$ can be understood in terms of surface frontogenesis. Let us consider a front undergoing frontogenesis as represented in Fig. 3. In the following reasoning we assume that $\Phi^0$ has zero horizontal average at the surface because $\Phi^0$ is a streamfunction anomaly. Since $\partial \Phi^0 = -\rho^0$ and velocities are vanishing at depth, we shall have at the surface $\Phi^0 > 0$ (resp., $\Phi^0 < 0$) on the warm (resp., cold) side. As for the ageostrophic secondary circulation, it is made of an ascending branch on the warm side and a descending branch on the cold side (Hoskins & Bretherton 1972). The condition $w = 0$ at the surface yields $\Phi^0 w_2^1 < 0$ on both sides of
the front. Since frontogenesis statistically dominates over frontolysis in the wavenumber range where the forward cascade of surface density variance is taking place one expects \(-\hat{w}_1\hat{\Phi}^*\) to be positive there, which is consistent with the shape of \(\Pi_a\) (Fig. 2a).

A more quantitative analysis allows one to relate \(\Pi_a\) to the release of available potential energy by the frontogenesis mechanisms. This is done by using (2.6), (2.10) and the thermal wind balance. The thermal wind balance, deduced from the geostrophic and hydrostatic approximations (2.1), is

\[
\nabla_H \rho = k \wedge \partial_z u^0 . \quad (4.1)
\]

Taking \(\nabla_H\) of (2.6), subtracting \(k \wedge\) of (2.10) (which is valid at any depth), and using the thermal wind balance, yields the relation between the frontogenetic vector \(Q\) and the ageostrophic flow (Hoskins & Bretherton 1972; Klein et al. 1998)

\[
2Q = \nabla_H w^1 - u^a g . \quad (4.2)
\]

with \(Q = [\nabla_H u^0]^T . \nabla_H \rho^0 . \) \([\ ]^T\) designates the transpose matrix. (4.2), that comprises the QG version of the Eliassen-Sawyer equation (Thomas & Lee 2005), leads to

\[
\frac{2}{K} \hat{Q} \cdot \hat{\nabla}_H \rho^0 * = K \hat{w}_1 \hat{\rho}^0 * - \frac{1}{K} \hat{u}_z^a g \cdot \hat{\nabla}_H \rho^0 * . \quad (4.3)
\]

However, (4.3) can also be written as (using \(\nabla_H \cdot u^a g = -w_1^1\))

\[
\frac{2}{K} \hat{Q} \cdot \hat{\nabla}_H \rho^0 * = [\hat{w}_1^1 \hat{\Phi}^0 * - \hat{w}_1^1 \hat{\Phi}^0 *]_z . \quad (4.4)
\]

Integrating (4.4) over the whole water column (with \(w_1^1 = 0\) at the boundaries and assuming \(\Phi^0 = 0\) at the bottom) and then with respect to \(K\) yields

\[
- \int_K^{\infty} < \text{Re} \left[ \frac{2}{K} \int_{-\infty}^{0} \hat{Q} \cdot \hat{\nabla}_H \rho^0 * \, dz \right] > dK = - \int_K^{\infty} < \text{Re} \left[ w_1^1 \hat{\Phi}^0 * \bigg|_{z=0} \right] > dK = \Pi_a ,
\]

which was verified in our numerical solutions. The same integration but involving (4.3) yields

\[
\Pi_a = \int_K^{\infty} < \text{Re} \left[ -K \int_{-\infty}^{0} \hat{w}_1 \hat{\rho}^0 * \, dz + \frac{1}{K} \int_{-\infty}^{0} \hat{u}_z^a g \cdot \hat{\nabla}_H \rho^0 * \, dz \right] > dK . \quad (4.5)
\]
(4.5) indicates that $\Pi_a$ is directly related to the production of density gradients within the fluid. This term should be positive in particular in the high wavenumber region, where density gradients are the strongest (since density gradient spectrum has a $k^{1/3}$ slope). (4.6) further indicates that $\Pi_a$ is related to the release of available potential energy by the ageostrophic circulation associated with frontogenesis. Indeed, the first term on the right hand side of (4.6) corresponds to a buoyancy flux while the second term reflects the manner by which ageostrophic flows modify the stratification through differential horizontal advection. For frontogenetic conditions (see Fig. 3), both of these terms are positive and lead to a reduction of the available potential energy. Thus, $\Pi_a$, directly related to the production of density gradients within the fluid, can be interpreted as a transformation of available potential energy into surface KE. From (2.16), the consequence of the positiveness of both $\Pi_a$ (related to the frontogenesis strength) and $\Pi_p$ (related to the direct cascade of density variance) and of the dominance of $\Pi_a$ is that $\Pi_a$ (related to the nonlinear advective transfers of surface KE) must be negative over most of the spectral range. These findings emphasize the importance of small-scale frontogenesis in the surface KE budget (as also noted in Cape et al. (2007) and Klein et al. (2007)).

5. Conclusion

We have revisited the spectral transfer properties of the SQG theory. Although the velocity and density spectra at the surface are identical, their spectral budgets differ. Surface density variance experiences a forward cascade as stated by Blumen (1978) and confirmed by numerous subsequent studies. Surface KE on the other hand experiences a clear and significant inverse cascade on a large range of scales, i.e., a transfer through advection by the geostrophic flow from small to larger scales. This apparent contradiction is explained by the presence in the surface KE budget of an additional term associated with the ageostrophic component of the flow, that is understood as a conversion of available potential energy into KE. Indeed its main underlying physics is the
restoration of the thermal wind balance, tending to be destroyed by the frontogenesis processes near the surface.

In systems more complex than SQG, surface density and velocity spectra have also been found to remarkably coincide, at least over a large part of the submesoscale and mesoscale range as shown in the PE framework by Klein et al. (2007) (see Fig. 4a that reproduces their spectra; their simulations are for a baroclinically unstable quasi-equilibrated jet in a periodic channel). But again the similarity of these surface spectra does not imply a similarity of their respective advective fluxes. In fact, and as found in SQG, surface density variance is fluxed downscale whereas KE is fluxed upscale over a wide central part of the wavenumber range (Fig.4b). The KE advective flux is compensated for by an ageostrophic term whose expression is similar to \( \Pi_a \). The main difference with our SQG results is that nonlinear interactions involve advection of the geostrophic quantities not only by the geostrophic velocities but also by the ageostrophic velocities. As a result, both \( \Pi_a \) and \( \Pi_u \) give a net contribution to the surface KE budget for PE solutions. These results corroborate previous evidences of a strong impact of the ageostrophic circulation on near-surface KE spectral transfers (Capet et al. 2007; Klein et al. 2007).

Overall the comparison between SQG and PE energy transfers suggest SQG as a pertinent framework to understand the inverse KE cascade found by Scott & Wang (2005) using altimetry data. A rapidly growing body of numerical (Lapeyre & Klein 2006a; Klein et al. 2007) and observational (LaCasce & Mahadevan 2006; Isern-Fontanet et al. 2006; Le Traon et al. 2007) evidences indicate that the ocean sea level does significantly reflect surface modes (as opposed to primarily the interior first baroclinic and barotropic modes Scott & Arbic 2007) even for scales above the deformation radius, which further strengthens this view.

One weakness of SQG is that it exhibits a downscale surface KE flux for high wavenumbers that is weak and resolution dependent (it is basically confined to the dissipation range). On the other hand high resolution PE (Capet et al. 2007) or Boussinesq (Molemaker et al. 2007) solutions produce robust downscale surface KE within the small-scale range (\(< 20-30km\)). This discrepancy can be traced to momentum advection by the ageostrophic flow as shown in Fig.4b, where the advective KE flux associated with the total flow and with its geostrophic component only can be compared (see also Capet et al. 2007). Molemaker et al. (2007) further relate the non-QG effects to ageostrophic frontal instabilities at submesoscale. This downscale flux of KE at submesoscale inadequately reproduced within the QG approximations may play an essential role in the dissipation of oceanic energy.

This work is supported by the CNRS and IFREMER (FRANCE). PK and BLH also acknowledge the support from the French ANR (Agence Nationale pour la Recherche, Contract no ANR-05-CIGC-010). We thank two anonymous reviewers for their thoughtful comments on the manuscript, which led to improve the discussion in section 4.

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