A Quasigeostrophic Analysis of a Meander in the Palamós Canyon: Vertical Velocity, Geopotential Tendency, and a Relocation Technique

ANANDA PASCUAL* AND DAMIÀ GOMIS
Institut Mediterrani d’Estudis Avançats, IMEDEA (CSIC-UIB), Mallorca, Spain

ROBERT L. HANEY
Department of Meteorology, Naval Postgraduate School, Monterey, California

SIMÓN RUIZ
Institut Mediterrani d’Estudis Avançats, IMEDEA (CSIC-UIB), Mallorca, Spain

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ABSTRACT

The quasigeostrophic geopotential tendency and omega equations are integrated to examine the dynamics of an upper-ocean coastal meander sampled during an intensive survey in the Palamós Canyon (northwestern Mediterranean Sea). Results for dynamic height tendency reveal that the meander is not growing or decaying but is propagating downstream at a speed of about 4 km day\(^{-1}\). This propagation speed implies a problem of synopticity in the field observations, which is solved through a data relocation scheme. The station relocation has important consequences on the magnitude of crucial dynamical variables such as the vertical velocity: maximum values of 20 m day\(^{-1}\) before the relocation reduce to 10 m day\(^{-1}\) after the relocation. The impact of bottom boundary conditions in the solution of the omega equation is also analyzed. Results indicate that for the stratification encountered during the survey, effects of topographic forcing are negligible above approximately 300 m.

1. Introduction

In the northwestern Mediterranean Sea, the “Northern Current” (Millot 1999) flows in a cyclonic sense along the Italian, French, and Spanish coasts (Fig. 1). Previous observations on the northeast Spanish shelf (Masó et al. 1990; Tintore et al. 1990; Alvarez et al. 1996) have shown some evidences of flow modifications (meanders, eddies, shelf–slope exchanges) that have been tentatively related to the presence of abrupt canyon topography. However, most of the existing samplings were too coarse for an accurate inference of crucial dynamical variables. The vertical velocity, for instance, cannot be measured directly, and usually has been computed through the quasigeostrophic (QG) omega equation (Tintore et al. 1991; Pollard and Regier 1992; Allen and Smeed 1996; Pinot et al. 1996). For a correct estimate of the QG vertical velocity it is highly recommended to have a regular and dense spatial sampling (Allen et al. 2001; Gomis et al. 2001). However, this will always be achieved at the expense of a loss of synopticity, so that an optimal compromise between spatial resolution and synopticity must be determined (Allen et al. 2001).

The lack of synopticity has already been addressed by Rixen et al. (2001, 2003) and Allen et al. (2001). The first proposed a correction method based on a geostrophic relocation of the stations, demonstrating that it can lead to some improvement in the accuracy of vertical velocities diagnosed from nonsynoptic observations. A drawback of this method is that it is based on flow advection instead of on pattern advection. Hence, for stationary fields there is no need to relocate observations of dynamical variables even when they are not synoptic, whereas Rixen et al.’s method will distort the original sampling. (Nevertheless, the method could still be valuable for the mapping of bio–geochemical variables advected by the geostrophic flow.) Allen et al. (2001) found a simple analytical expression relating the asynopticity of a survey to the mean flow, to the propagation velocity of the perturbation and to the speed of the ship in the direction of the propagation. However,
the propagation velocity is assumed to be known, and therefore no guidance is provided on how it could be computed from survey data. More recently, Rixen et al. (2003) have tested along- and cross-front survey strategies with the aim of determining which option minimizes the errors associated with the synopticity of the observations. Their results suggest that an along-front strategy will produce better results when the front is significantly distorted by propagating instabilities, whereas a cross-front sampling will be more accurate in the absence of significant frontal distortion.

In this paper we present results from an intensive survey in the Palamós Canyon. The sampling strategy consisted of a regular station distribution, dense enough to compute QG vertical velocities at the mesoscale. The second condition (synopticity) will be studied by estimating the temporal variability of the fields from the QG geopotential tendency equation. Namely, we present a technique to determine the propagation velocity of a structure and to relocate the stations in order to minimize the lack of synopticity. A first aim of this study is to investigate the impact of the lack of synopticity and of the proposed station relocation on the computation of dynamical variables from survey data.

The ultimate goal is to study the QG dynamics of the sampled meander, including the eventual impact of bottom topography on vertical velocity estimates. Part of the difficulty is due to the fact that both the internal Rossby radius and the scale of the canyon topography are \( O(15 \text{ km}) \). Since the length scale of the main structure sampled in this study is also \( O(15 \text{ km}) \) (about 60-km wavelength), this implies that both baroclinic instability (local or remote) and canyon topographic forcing must in principle be considered as potential mechanisms for the formation of the meander.

2. Data analysis

a. The dataset and data interpolation

The data used in this study come from the intensive oceanographic cruise CANONS II, carried out between 24 and 31 May 2001. The main objective of the survey was to investigate the hydrodynamics induced by the Palamós Canyon. A domain of \( 80 \times 70 \text{ km}^2 \) was covered by 134 CTD stations separated 4 km just over the canyon (a subdomain of \( 25 \times 40 \text{ km}^2 \)) and 8 km elsewhere (Fig. 2). CTD casts reached a maximum depth of 600 m, which is considered a reliable reference level in the region (e.g., Pinot et al. 1995).

Direct velocity measurements were collected using a 153.6-kHz vessel-mounted acoustic Doppler current profiler (VM-ADCP). The vertical resolution was configured to 8 m and the sampling period to 2 min. Major error sources for ADCP measurements (i.e., gyrocompass error and ship's velocity estimation) were reduced using the methodology described by Griffiths (1994) and Allen et al. (1997). After calibration and correction of errors induced by ship motion, individual profiles were averaged over 10-min ensembles. The cruise took place under excellent weather conditions, with no wind on the previous days nor during the entire week of the cruise.
The station separation (and not on the size of the struc-
turing. The degree of smoothing therefore depends on

to filter out scales that cannot be resolved by the sam-

applied in the way proposed by Pedder (1993) in order

on the size of the dominant structures of the corre-

lation scale was set to 10 km according to correlation

(see Fig. 3). This parameter depends basically

Regarding objective analysis parameters, the corre-

scale for the objective analysis scheme. Similar results

were obtained when applied to other levels.

Therefore, no wind-driven circulation features such as
coastal upwelling or inertial waves are expected. Tidal

Dashed lines are three Gaussian functions with widths of 5, 10, and 15 km.
The dashed line (L = 10 km) gives the best fit and is therefore taken

as correlation scale for the objective analysis scheme. Similar results

were obtained when applied to other levels.

The measured hydrodynamical variables were inter-

polated from observation points onto grid points using

a successive correction scheme with weights normalized

in the observation space in order to approach optimum

statistical interpolation (Bratseth 1986). The domain

was the same for both the gridding and diagnostic cal-

culations. In the vertical direction it consisted of 60

equally spaced levels spanning from 10 to 600 m. The

horizontal domain was a grid of 37 \times 42 points sepa-

rated 2 km in both the offshore (x) and alongshore (y)
directions (the 74 \times 84 km² area is rotated 27° clockwise

with respect to longitude/latitude axis; see Fig. 2).

Regarding objective analysis parameters, the corre-

lation scale was set to 10 km according to correlation

statistics (see Fig. 3). This parameter depends basically

on the size of the dominant structures of the corre-

sponse variable (and not on how they are sampled).

An additional normal-error filter convolution was also

applied in the way proposed by Pedder (1993) in order
to filter out scales that cannot be resolved by the sam-

pling. The degree of smoothing therefore depends on

the station separation (and not on the size of the struc-
tures being sampled). This process is especially crucial
when nonlinear combinations of high-order derived var-

iables (e.g., vorticity advection) are to be obtained from

the interpolated fields. Otherwise, derived fields are

completely dominated by the poorly resolved small

scales. In our case the cutoff wavelength was set to 32

km [4 times the maximum station separation, as rec-

ommended by Gomis et al. (2001)].

The computation of dynamic height in the presence

topography deserves some further attention. Our ap-

proach consisted of not computing dynamic height re-

ferred to 600 m prior to its interpolation. Instead, the

variable computed at station points and interpolated onto

grid points was layer thickness (equivalent to referring
dynamic height to the level immediately below). After

interpolation, dynamic height was recovered at each grid

division simply as the sum of all layer thickness values
down to the reference level. Hence, at points shallower

than 600 m the underground water column is inferred

from the nearest stations and added to the existing piece

of water column. This is approximately equivalent to

integrating specific volume anomaly along the slope, as

suggested by Csanady (1979). An indirect proof for the

reliability of the described method is that dynamic

height patterns do not show any apparent discontinuity

across the 600-m isobath.

Derived dynamical variables such as geostrophic ve-

locity or geostrophic vorticity were simply obtained by

finite differences from interpolated grid point values of
density and dynamic height. The computation of vertical

velocities and geopotential tendency are more cumber-
some and are therefore described in more detail in the

next section.

b. The quasigeostrophic omega equation

Vertical velocities \( w \) were obtained by integrating the

QG omega equation on an \( f \) plane (Holton 1992):

\[
\left( N^2 \nabla^2 + f_0 \frac{\partial^2}{\partial z^2} \right) w = f_0 \frac{\partial}{\partial z} (v_y \cdot \nabla \zeta) + \frac{g}{\rho_0} \nabla \cdot (v_y \cdot \nabla \rho),
\]

where constants \( g \) and \( f_0 \) are the gravitational acceleration

and the Coriolis parameter at the domain latitude; vari-

able \( \rho \) is the difference between the total density and a

constant reference density \( \rho_o \); \( v_y \) is the geostrophic ve-

locity; \( \zeta \) is the vertical component of geostrophic relative

vorticity, defined as \( \zeta_y = f_0 \nabla \Phi \) where \( \Phi \) is the geo-
potential or dynamic height and subscript \( h \) denotes hori-
zontal differentiation; and \( N \) is the Brunt–Väisälä fre-
cquency defined as \( N^2 = -\frac{g}{\rho_0} \frac{\partial \rho}{\partial z} \), where \( \rho \) is the

mean-level density (i.e., \( N^2 \) is assumed to have only ver-
tical dependence).

Initially, the equation was integrated setting \( w = 0 \)
at the upper and lower boundaries (10 and 600 m) and

setting its normal derivative to zero at the lateral bound-

aries (Neumann conditions). [For a detailed description

\[ \text{Unauthenticated} \mid \text{Downloaded 09/18/23 08:44 PM UTC} \]
of the method see Pinot et al. (1996).] It is worth mentioning here that the differences between setting the integrated variable to zero at 600 m (despite part of the domain being underground) or at the actual bottom depth are quite small. A first reason is that the strongest part of the forcing is located offshore with respect to the location of the 600-m isobath. The second is that the underground grid values mostly consist of a linear trend as soon as they depart a distance greater than the correlation scale from the nearest station. A linear dynamic height field alone does not contribute to the omega forcing terms. (The same will hold true for the tendency equation.)

While for the upper and lateral boundaries the assumptions seem reasonable, this is not necessarily the case for the lower boundary where vertical velocities could be different from zero if the submarine canyon was inducing a significant upsloping or downsloping velocity. As a first approximation to study the influence of the bottom topography we will compute the vertical velocity at the bottom boundary of the domain using the approximate kinematic boundary condition: \( w_b = -v_h \cdot \nabla d \), where \( d \) is the bathymetry depth (see below).

c. The quasigeostrophic tendency equation

In the framework of QG dynamics it is also possible to find an equation that relates the geopotential \( \Phi \) and \( \partial \Phi / \partial t \) without explicitly determining the distribution of \( w \) (Holton 1992, p. 158). Defining the geopotential tendency as \( \chi = \partial \Phi / \partial t \), the tendency equation for the ocean can be written as

\[
\frac{\partial \chi}{\partial z} = \frac{2}{\rho} \nabla \cdot (v \, \nabla \rho).
\]

The local time tendency of geopotential can be separated into two contributions: translation, on one hand, and change of shape and/or intensity, on the other hand. For waves, if maxima/minima of tendency were mostly located in between ridges/troughs, this would indicate that the wave is propagating but not growing/decaying. Instead, if maxima/minima were located over ridges/troughs, the wave would be mainly growing or decaying (Holton 1992). In the atmosphere, the translation contribution is reported to dominate over amplification or decay in most situations (Carlson 1991). In the ocean, the question is still open.

A time scale for the dominant features can be estimated as \( \Delta t = \chi^{-1} \Delta \Phi \). If the estimated time scale is much larger than the time required for sampling the domain, then observations can be considered synoptic. Conversely, a problem of lack of synopticity arises when the sampled structure is evolving with a time scale comparable to the sampling time. Because this will be the situation for this study, we will first address the problem of synopticity in the case of a disturbance that is propagating at a velocity not much less than the sampling speed.

3. Relocation method

The method, developed with the aim of reducing the problem of synopticity, consists of an iterative procedure. Each iteration can be split into two different steps: 1) computing a system phase velocity and 2) relocating the CTD stations according to the system speed obtained in the previous step.

To obtain a system phase velocity we recall the Eulerian and Lagrangian system formulations. In the Eulerian system, the change of a scalar variable with respect to time is expressed as a local derivative. In the Lagrangian system, instead, the change of a scalar variable following a material parcel is expressed as a total derivative consisting of a local term and an advective term. For the particular case of the geopotential \( \Phi \),

\[
\frac{D\Phi}{Dt} = \frac{\partial \Phi}{\partial t} + (v - c) \cdot \nabla \Phi.
\]

Now consider a frame of reference moving at a constant horizontal velocity \( c = (c_x, c_y) \). In this frame, the total derivative remains the same, whereas the terms on the right side of (3) change to

\[
\frac{D\Phi}{Dt} = \frac{\partial \Phi}{\partial t} \bigg|_c + (v - c) \cdot \nabla \Phi.
\]

By comparing (3) and (4), we obtain the relation between the local terms written in the fixed and the moving frames of reference:
FIG. 4. Mean Brunt–Väisälä frequency profile (s\(^{-1}\)) computed from the original density profiles reaching 600 m. The horizontal dashed line represents the interface between the two layers considered in Tang’s model; the vertical dashed lines correspond to the constant Brunt–Väisälä frequency assigned to each of the two layers. Also shown is an example of an \(N^2\) profile (station 77) located on the inshore side of the front.

\[
\frac{\partial \Phi}{\partial t} = \frac{\partial \Phi}{\partial t} \mid_A - \mathbf{c} \cdot \nabla \Phi \mid_B \tag{5}
\]

Term B corresponds to the propagation or advection term and A to the growing term in the moving frame of reference. A natural way of defining a system velocity is to associate it with the value of \(c\) that minimizes the magnitude of \(\partial \Phi / \partial t\), within the region of interest (e.g., as its minimum mean square; see appendix A for further details). Finding a value of \(c\) that makes \(\partial \Phi / \partial t\) exactly equal to zero everywhere would mean that the whole observed field is propagating exactly at that speed, with no growing/decaying.

Once the system speed is determined, the goal is to obtain a synoptic map of our data, that is, to infer the position of the stations at a common instant, such as the time corresponding to the midtime of the survey. To relocate every single observation to the position that it would occupy at the reference time, we can simply use the following transformation:

\[
X_i = x_i - c_i t_i \quad \text{and} \quad Y_i = y_i - c_i t_i, \tag{6}
\]

where \(t_i\) is the time of observations measured relative to the midtime of the survey, and \((x_i, y_i)\) are the known horizontal location of observations. This procedure obviously assumes that the phase speed can be considered as constant during the entire sampling period. Once the stations are relocated, we shall proceed to reanalyze the observed variables and infer the dynamically derived variables.

4. Results

We will focus on the fields at 200 m, well below the thermocline that was found at about 100-m depth (see Fig. 4). At a depth of 200 m the dominant signature is a shelf–slope salinity front (the Catalan front; Font et al. 1988) separating fresher coastal waters from the denser offshore waters. Figure 5 shows the raw (un-
gridded and unsmoothed) salinity and density fields, where the salinity front and a meander imbedded in it are well apparent. This figure also reinforces the need for some kind of filtering in order to eliminate small-scale features that cannot be resolved by the sampling. All of the following figures will show results obtained in the way described in the previous section (i.e., gridded, filtered fields).

a. The original dynamical fields

The geostrophic current associated with the shelf-slope front (the so-called Northern Current; Millot 1999) flows to the southwest (Fig. 6a). At the northern boundary of the domain the flow is deflected onshore with a clear anticyclonic curvature, whereas just upstream from and above the canyon it describes a cyclonic meander. Downstream of the canyon the flow veers again toward the coast with a slight anticyclonic vorticity. The meander has a wavelength of about 50–60 km, a cross-front amplitude of about 20 km, and maximum geostrophic velocities of 13 cm s$^{-1}$ (at 200 m). Using the scale values $U \sim 13$ cm s$^{-1}$, $f \sim 10^{-4}$ s$^{-1}$, and $L$ (quarter wavelength) $\sim 13$ km we obtain a value of about 0.1 for the Rossby number, which justifies the QG analysis presented below.

An additional evidence for the fulfillment of the geostrophic balance comes from the comparison between geostrophic and ADCP velocities (Fig. 7). In order to obtain differences between both fields, ADCP observations were interpolated onto the same grid and submitted to the same filter as the CTD data. Horizontally averaged rms differences between both fields were 6.41 cm s$^{-1}$ at 50 m, 6.45 cm s$^{-1}$ at 100 m, and 4.86 cm s$^{-1}$ at 200 m, and the spatial distribution of ADCP minus geostrophic velocity (not shown) is quite random. This
suggests that the differences may be related to ADCP instrumental errors and not to a subjacent large-scale (smooth) barotropic flow or to the existence of significant currents at or below the reference level. In summary, geostrophic velocities seem to provide an accurate approximation to absolute velocities.

The QG vertical velocity field (Fig. 6b) obtained using (1) shows maximum downward (upward) velocities where anticyclonic (cyclonic) vorticity is advected by the mean flow. This indicates that the vertically differential advection of relative vorticity by the mean flow [first term on the right-hand side of (1)] dominates over density advection (second term on the same side). In our domain this means that downward motion is obtained before the upstream wall and after the downstream wall of the canyon, whereas upward velocities are obtained above the canyon. Maximum values of vertical velocity are about 20 m day\(^{-1}\) (both positive and negative). Since the main vertical velocity structures are located offshore with respect to the 600-m isobath, they are not likely distorted by any artifact related to the estimation of dynamic height in the presence of a shallow bathymetry.

On the other hand, the distribution and magnitude of vertical velocities are subject to some uncertainties due to the boundary conditions assumed for the integration of the omega equation. In particular, the selection of a bottom boundary condition could be critical in our case if the submarine canyon were forcing a significant up- or downslope vertical motion at the deepest layer of our domain (600 m). To estimate this effect we have integrated the omega equation with the following lower boundary condition:

\[ w_b = -v_b \cdot \nabla d, \]

where \( w_b \) is the vertical velocity to be imposed at the lower boundary, \( d \) is the bathymetry depth (available on a grid of 1-km resolution from a digitalization of a nautical chart), and \( v_b \) is the horizontal geostrophic velocity at 590 m (the first level above the reference level). Equation (7) is a linear approximation to the kinematic lower boundary condition of no flow through the seafloor. This calculation, using the approximate condition (7), is only intended to obtain a general estimate of the bathymetric effect and its influence on the vertical velocities in the interior. As we will see, the effect of the bottom depends more on the vertical stratification than it does on the value of \( w_b \) itself. This approximate lower boundary condition was first justified for the atmosphere by Phillips (1963) and was subsequently used many times to diagnose the effect of topography on atmospheric flows (e.g., Krishnamurti 1968; Derome and Winn-Nielsen 1971; Egger 1976). The result (Fig. 8) is a reasonable field of vertical velocities at the bottom of the domain: sinking motion at the northern (upstream) wall of the canyon and rising motion at the southern (downstream) wall with a magnitude of about 10–15 m day\(^{-1}\). This diagnostic estimate is very similar to the result obtained by Klink (1996) in a similar “downwelling” canyon simulation using a complete numerical primitive equation model.

Figure 9 compares, in a vertical section perpendicular to the canyon axis, the vertical velocities obtained from integrating the omega equation with the two sets of lower boundary conditions [(a) \( w_b = 0 \), and (b) \( w_b = -v_b \cdot \nabla d \)]. The figure reveals that the main structures of the bottom-induced vertical velocity and of the QG vertical velocities induced by vorticity advection in the meander are displaced relative to each other. However, the most revealing result is that the effect of the bottom on the vertical velocity vanishes at and above about 300 m, being completely negligible at the level on which we are focusing (200 m). In appendix B it is shown that only for a wider canyon and for weaker stratification conditions could the topographic effect be important over the whole water column.

b. Testing the synopticity of observations

Further evidence for the interaction between the meander and the canyon can be obtained from the study of the temporal evolution of the meander. In order to determine if the meander is growing/decaying and/or propagating, we make use of the tendency equation. Figure 10 presents the geopotential tendency at 200 m, computed from (2). First of all, it is worth noting the position of the maxima/minima with respect to the ridges and troughs depicted in Fig. 6a. Positive (negative) values of tendency mean that dynamic height is increasing (decreasing). In our case, the positive (negative) centers are located downstream of the meander ridges
Fig. 9. Vertical section of vertical velocity obtained from integrating the omega equation with two different lower boundary conditions: (a) \( w_b = 0 \) and (b) \( w_b = -v_b \cdot \nabla d \). The section is 40 km long perpendicular to the canyon axis. The direction of the mean flow is from the right to left in the figure.

The propagation of the meander raises a potentially important problem of synopticity. As noted above, the value of \( c \) that minimizes \( \partial \phi / \partial t \bigg|_c \) in (5) over the domain of interest provides an estimate of the propagation velocity. For this purpose, we selected a domain where the largest tendencies are located: between 40 and 70 km in the alongshore direction, between 30 and 70 km in the offshore direction, and between 100 and 250 m in the vertical direction. The result of the minimization (see appendix A for details) was \( c_x = 0.054 \text{ km day}^{-1} \) and \( c_y = -6.168 \text{ km day}^{-1} \), where the \( x \) and \( y \) directions refer to the cross-shore (positive offshore) and along-shore (positive with shore to the left) directions, respectively. As expected, the value of \( c_x \) is negligible but \( c_y \) is quite important, especially when compared with the sampling velocity. The field survey was carried out upstream (from south to north) in 6.6 days and the length of the domain is 72 km, which yields an average upstream sampling velocity of about \( \sim 11 \text{ km day}^{-1} \).

c. Station relocation and final fields

In an attempt to reduce the synopticity problem we relocated the position of each CTD station based on the linear transformation given in (6), where the time of reference was set to the middle time of the cruise (i.e., day 3.3). The system velocity used to relocate the stations was the one obtained in the previous subsection (rounding off, \( c_x = 0 \text{ km day}^{-1} \) and \( c_y = -6 \text{ km day}^{-1} \)).

Once the station data were relocated, all variables were reanalyzed. To be consistent, the analysis parameters (the correlation scale and cutoff wavelength) were recalculated. After computing the correlation statistics, the characteristic scale did not show a significant change (not shown), and so it was kept equal to 10 km. How-
ever, the cutoff wavelength had to be modified since the separation between stations changed because of the relocation. The new maximum station separation in the relocated direction that is, alongshore, can be computed as $dy = (1 - c_x/v)dy_o$ (see Allen et al. 2001), where $dy_o$ is the original maximum distance (8 km) and $v$ is the velocity of the ship in the alongshore direction (11 km day$^{-1}$). With the value of $c_x = -6$ km day$^{-1}$ it yields a new station separation distance of 12.36 km and consequently a new cutoff wavelength of $\sim 50$ km.

From the new dynamic height analysis we recomputed all derived variables, including the vertical velocity and the dynamic height tendency (not shown). The latter reveals again a meander propagating downstream at a certain speed, which we can recalculate through the minimization of (5). The results obtained for the propagation speed were $c_x = -0.150$ km day$^{-1}$ and $c_y = -4.128$ km day$^{-1}$. It is not surprising that the values of the propagation speed did not coincide with the first computation because the original fields were analyzed without taking the moving system into account. In practice an iterative procedure is necessary to finally determine the propagation speed. For each iteration all fields must be relocated with the value of the system velocity obtained in the previous iteration until the procedure converges. In our case, after the second relocation we obtained a new set of system velocities: $c_x = -0.140$ km day$^{-1}$ and $c_y = -4.688$ km day$^{-1}$. Hence, two iterations were apparently enough for the computations to converge. Final results show a meander propagating downstream at a velocity of $\sim 4$ km day$^{-1}$.

Figure 11 presents the final fields, after the relocation derived from a phase speed of 4 km day$^{-1}$. While for dynamic height and geostrophic velocity the differences between original and relocated fields reflect only a stretching of the meander, changes are more significant for high-order derived variables such as the vertical velocity. The magnitude of vertical velocities is actually reduced to about a half of the original values (Fig. 11b), maximum values now being about $\pm 10$ m day$^{-1}$. The reason is that, for an upstream survey (with respect to the propagation of structures), wavelengths along the sampling direction are underestimated. Hence, a first analysis will overestimate the derivative terms in that direction and, consequently, will also overestimate the vertical forcing and the vertical motion itself.

A relevant question is to what extent expression (5) is minimized with the value of $c_x = -4$ km day$^{-1}$. For an idealized wave propagating at a single and well-defined velocity without changing its shape (i.e., a neutral mode), $\partial \Phi/\partial t$ would be exactly zero and then the minimization would be perfect. In the real ocean we normally find a combination of modes propagating at different system velocities, with some of them growing and some decaying. In this case, the method proposed here will not be exact, but will nevertheless improve the original field. A natural way of measuring the extent of the minimization is to compare the variance of the tendency in the fixed frame of reference $(\partial \Phi/\partial t)$ and in the moving frame of reference $(\partial \Phi/\partial t_i)$. In our case, the standard deviation (over the region indicated above) was 1.625 $10^{-6}$ dyn cm$^{-1}$ for the former and 0.334 $10^{-6}$ dyn cm$^{-1}$ for the latter. We can therefore state that about 96% of the variance of the original tendency field was due to the downstream propagation of the meander at a speed of 4 km day$^{-1}$. The remaining 4% could be due to the contribution of other neutral modes or also to some growing modes, but in any case they account for a negligible percentage of the variability.

5. Discussion and conclusions

a. On the reliability of results

A first key point to be discussed is the adequacy of examining a meander flowing over a canyon in the framework of QG theory. The geostrophic velocity has been shown to be in good agreement with ADCP observations and the Rossby number is of $O(0.1)$, which in principle confirms the suitability of QG theory at least over most of the domain (except at the bottom boundary layer). The impact of different bottom boundary conditions on the solution of the omega equation has also been evaluated. Results indicate that, for the typical stratification of the study region, the effects of topographic forcing are primarily constrained below the front structure so that this would not be significantly distorted by the canyon. Hence, the obtained vertical velocity pattern (downwelling upstream and along the upstream wall of the canyon and upwelling along the canyon axis) would be strictly related to the location of the meander at the cruise time and not to the canyon itself. The effect of friction could be included in the diagnostic calculations by introducing also an Ekman vertical velocity in the lower boundary condition in the same way that the kinematic effect is introduced. The Ekman value would be estimated as

$$w_E = |w(2f)|^{1/2} \xi_{s}$$

(8)

[see, e.g., Gill 1982, p. 331, their (9.6.5)]. With $v = 0.003$ m$^2$ s$^{-1}$ (Gill 1982, p. 332), and $\xi_{s} \sim 10^{-6}$ s$^{-1}$ at 560 m, we get $w_E \sim 0.4$ m day$^{-1}$. This is entirely negligible in comparison with the QG values of $w$ in the interior, as well as the estimates of $w_E$ at the bottom.

Perhaps the most innovative aspect of this work has been use of the geopotential tendency equation (Holton 1992), an equation traditionally used in meteorological studies. Here it has been integrated in order to examine the temporal variability of the fields. Results have revealed that the meander is not growing or decaying, but basically propagating downstream at a velocity of about 4 km day$^{-1}$. Such a propagation has raised a problem of synopticity in the field observations, which has been addressed through a data relocation scheme. The station relocation has shown important consequences on the
Fig. 11. Horizontal maps at 200 m after a relocation derived from a phase speed $c_y = 4$ km day$^{-1}$ ($c_x = 0$): (a) dynamic height with vectors of geostrophic velocity over-imposed, (b) vertical velocity, and (c) tendency ($10^{-6}$ dyn cm s$^{-1}$).
magnitude of the vertical velocities, which have reduced from \(-20\) to \(-10\) m day\(^{-1}\) after the station relocation.

Errors in the computation of vertical velocities due to nonsynoptic surveys have already been examined by Allen et al. (2001) with both a numerical and an analytical model. The analytical model helps to understand the different components of the errors and indicates key parameters for the design of mesoscale sampling. For a comparison with our results, we can use the expression obtained by Allen et al. (2001), which relates the apparent (nonsynoptic) vertical velocity \(w_a\) to the actual (synoptic) vertical velocity \(w\):

\[
w_a = w(1 - K)e^{\gamma},
\]

\[
K = \frac{c_r}{v_s},
\]

\[
\gamma = -\frac{c_r}{v_s}, \quad \text{and}
\]

\[
A = 1 + \frac{2k^2}{(k^2 + l^2)k^2 + l^2 + 1},
\]

where \(R_o\) is the internal deformation radius \((R_o = NH/f = 15\) km\), \(N = 5 \times 10^{-3}\) s\(^{-1}\) is the Brunt–Väisälä frequency over 300 m (see Fig. 4), \(k\) is the wavenumber in the alongshore propagation direction \((k = 2\pi/\lambda\) being the meander wavelength \(\lambda\) (70 km after the relocation; thus in our case \(k \approx 9 \times 10^{-5} \) m\(^{-1}\)), \(l\) is the wavenumber in the cross-propagation offshore direction (in practice it is 2 times the amplitude of the perturbation in the offshore direction, which can be taken as \(\approx 60\) km), and \(\gamma\) will be neglected since the meander is apparently propagating but not growing/decaying. Taking \(c_r = -4\) km day\(^{-1}\) and \(v_s = 11\) km day\(^{-1}\), (9) yields

\[
w_a = w(1 + 0.43).
\]

For the apparent vertical velocities \(w_a \approx 20\) m day\(^{-1}\) found in the present study, the actual (synoptic) vertical velocity \(w\) should be about 14 m day\(^{-1}\) (30% less than the apparent velocities). With our relocation technique we found maximum values of 10 m day\(^{-1}\); that is, the reduction of the magnitude was of about 50% between the apparent and relocated fields. Thus, our results overestimate the reduction of vertical velocity predicted by the theoretical model. Nevertheless, this disagreement can be explained because the theoretical model is a simplified model that assumes only two layers and a single linear instability mode. Hence, it assumes that we are resolving the same spectral contents both in the synoptic and nonsynoptic surveys. In practice, the horizontal resolution of the synoptic sampling is changed after relocation, and therefore the filtering has to be applied according to the new station separation. The increase in the cutoff wavelength of the filtering implies, by itself, a further reduction in the magnitude of vertical velocities.

In addition to the lack of synopticity, the magnitude of the vertical velocity is also affected by analysis errors derived from both errors in the observations and representation errors (i.e., those derived from the discrete sampling of a continuous field). Gomis and Pedder 2004, manuscript submitted to J. Mar. Syst.) have recently estimated the errors associated with the computation of vertical velocities from an oceanographic cruise with an aspect ratio between correlation scale and station separation similar to that of the present study. They obtained errors on the order of 15%–20% of the standard deviation of the vertical velocity field in the inner domain and of about 30% near the boundaries. [Consistent results (though for other derived variables such as vorticity) were found by Bretherton et al. (1976) and McWilliams et al. (1986).] This would indicate that, in the case of a meander like the observed one, the lack of synopticity would be the main error source if no station relocation is applied.

### b. On the dynamics of the meander

The ultimate aim of the cruise was to study the QG dynamics of the sampled meander. Part of the difficulty of the problem is due to the fact that both the internal Rossby radius, \(NH/f\) (Table 1), and the curvature radius of the canyon topography are \(O(15\) km\). Since the length scale of the meander is also \(O(15\) km\), this implies that baroclinic instability (local or remote) and topographic forcing should be both considered when attempting to explain the origin of the meander. Previous studies provide contradictory evidence: on the one hand, similar meanders observed during other hydrographic surveys in the northwestern Mediterranean have been tentatively related to the presence of submarine canyons (Masó et al. 1990; Tintoré et al. 1990; Alvarez et al. 1996). However, practically none of the samplings was dense enough for an accurate estimate of derived variables. Also some numerical models applied to submarine canyons support the idea of meanders trapped to the bottom topography (Klinck 1996; Ardhuin et al. 1999; Skliris et al. 2001, 2002) with little flow disruption in the upper levels above the canyon.

In the case of a buoyancy-driven coastal jet such as the present one, advective trapping can provide a par-

### Table 1. Input parameters for Tang’s model. The upper layer ranges from the surface to 300-m depth (Fig. 4), and velocity at surface \(U_s\) is 30 cm s\(^{-1}\). The lower layer is defined between 300 and 600 m and is considered a quiescent layer according to Tang’s model (in practice, velocities at 300 m are lower than 3 cm s\(^{-1}\)). The mean Brunt–Väisälä frequency values for each layer are obtained from Fig. 4.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Numerical value</th>
</tr>
</thead>
<tbody>
<tr>
<td>(U_s)</td>
<td>30 cm s(^{-1})</td>
</tr>
<tr>
<td>(H_1)</td>
<td>300 m</td>
</tr>
<tr>
<td>(N_1)</td>
<td>(5 \times 10^{-3}) s(^{-1})</td>
</tr>
<tr>
<td>(H_2)</td>
<td>300 m</td>
</tr>
<tr>
<td>(N_2)</td>
<td>(8 \times 10^{-4}) s(^{-1})</td>
</tr>
</tbody>
</table>
ticularly relevant mechanism for coupling the upper-ocean front to the bottom topography through a strong interaction between the front and the bottom boundary layer (Chapman and Lentz 1994; Chapman 2003). We do not include this potentially important mechanism in the present study because we ignore the bottom boundary layer and treat the dynamic effect of the bottom topography with a linear approximation. However, we do not consider this omission to be a critical deficiency. Our general impression is that the present coastal meander over the Palamós Canyon is in water that is deeper and generally more stratified than the typical shelf regions to which the advective trapping mechanism is most applicable.

At the same time, SST images of the Gulf of Lions reveal the presence of meanders of similar characteristics (wavelengths of about 80 km) propagating westward with velocities between 2.0 and 7.7 km day$^{-1}$ and then continuing southward along the Catalan coast (Flexas et al. 2002). Current-meter observations in the Gulf of Lions also support such a remote origin for this kind of meander, as they have provided evidence of flow modifications with periodicities in the range of 3–6 and 10–20 days (Sammari et al. 1995; Flexas et al. 2002). These authors have suggested baroclinic instability as a possible energy source, based on a simple analytical model proposed by Tang (1975).

Tang’s model assumes quasigeostrophy, which is a valid framework for the Northern Current as noted earlier. The model consists of an upper layer (thickness $H_1$, Brunt–Väisälä frequency $N_1$, maximum speed $U_1$, and constant shear $U/H_1$) over a quiescent lower layer ($H_2$, $N_2$). The model provides the propagation speed of the modes, as well as the boundary separating short stable waves from long unstable waves (i.e., the stability cutoff wavelength). Considering input parameters consistent with our study (see Table 1) we obtained a cutoff wavelength of about 90 km, which indicates that the meander sampled in this study (70-km wavelength) is in the stable regime. Furthermore, according to the model results the propagation speed of a wave with a wavelength of 70 km would be around 5 km day$^{-1}$ (Fig. 12), which is very similar to what we have found by means of the tendency equation.

Therefore, Tang’s (1975) simple analytical model seems to support the results of our QG analysis in that the observed meander, despite being located above the Palamós Canyon, is simply a neutral baroclinic mode propagating downstream at a velocity of about 4–5 km day$^{-1}$ without any noticeable interaction with the bottom topography. However, because of the simplicity of the analytical model (e.g., the two-layer assumption) and the sensitivity of results with respect to the input parameters, further evidence would be desirable. Numerical modeling would be a highly valuable tool for further research exploring the interaction between meandering coastal flow such as the Northern Current and the canyon orography, including the bottom boundary layer.

FIG. 12. Tang model results for the input parameters given in Table 1: (a) phase speed (km day$^{-1}$) and (b) time required for the wave to double its amplitude (days). The cutoff wavelength that separates short stable waves from long unstable waves is indicated with a dashed–dotted line.

For instance, it could determine a realistic ageostrophic component of the circulation and a more realistic estimate of the topographic forcing $U_0$, than is provided by the QG analysis presented in this work.

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**APPENDIX A**

**Computation of a System Velocity ($c_x$, $c_y$)**

The computation of a system phase velocity is accomplished through the minimization of $\partial \Phi / \partial t$, in (5). We seek the minimum of the square value of $\partial \Phi / \partial t$, averaged over the region of interest, which is the same as minimizing

$$J(c) = \int \left( \frac{\partial \Phi}{\partial t} + c \cdot \nabla \Phi \right)^2 dV, \quad (A1)$$

where $V$ denotes volume and the integral is over the 3D domain of interest. From the standard calculus of variations the minimum of $J(c)$ is obtained by imposing

$$\frac{\partial J(c)}{\partial c_x} = 0 \quad \text{and} \quad \frac{\partial J(c)}{\partial c_y} = 0. \quad (A2)$$

Taking into account that $c \cdot \nabla \Phi = c_x (\partial \Phi / \partial x) + c_y (\partial \Phi / \partial y)$...
The omega equation is simply

\[ N^2 \nabla^2 z + f^2 \frac{\partial^2 w}{\partial z^2} = 0, \tag{B1} \]

expressing the vertical velocity field as \( w = w_y \sin(kx) \cos(ly) F(z) \), where \( k \) and \( l \) are in this case related to the canyon width \( D \) and length \( L \) \( (k = 2\pi/2D \) and \( l = 2\pi/2L) \) and \( F(z) \) is a function that contains the vertical dependence. Substituting this expression in (B1) yields

\[-(k^2 + l^2)F(z) + \frac{f^2}{N^2} \frac{d^2 F(z)}{dz^2} = 0. \tag{B2}\]

The solution to (B2) is

\[ F(z) = Ae^{Sc} + Be^{-Sc}, \tag{B3} \]

where

\[ S = \sqrt{\frac{(k^2 + F)N^2}{f^2}}. \]

Now, imposing that the effect of the bottom has to be zero at the sea surface \((z = 0)\) and that at the bottom \((z = -z_b)\) we must have \( w = w_y \sin(kx) \cos(ly) \), we can determine the unknowns \( A \) and \( B \). It finally results in

\[ F(z) = w_y e^{Sc} - e^{-Sc}. \tag{B4} \]

Solution (B4) consists of an exponential decay that depends on the width and length of the canyon, the Brunt–Väisälä frequency, and the Coriolis parameter. Figure B1 shows examples for different stratification conditions and different sizes of the canyon \( (f_o \) is taken equal to \( 10^{-4} \text{ s}^{-1} \)). Hence, under the same stratification conditions, the effect of the bottom for a narrow canyon are more confined to the deepest layers than for a wide canyon. On the other hand, for a given canyon scale, a weak stratification will have the effect of spreading the bottom effect vertically. The data in Palamós Canyon are closest to the dashed curve \((N^2 \sim 10^{-5} \text{ s}^{-1})\) in the panel marked \( L = 15 \text{ km} \) (where \( L \) represents the horizontal scale of the canyon), for which the vertical scale is \( S^{-1} \sim 107 \text{ m} \) and the topographic effect is almost negligible at and above 300 m.

It is worth noting that \( N \) has been assumed to be constant, while in the real ocean it presents large gradients in the vertical direction (see Fig. 4). However, this simplified theoretical approach compares quite well to the results obtained from integrating the omega equation taking into account the vertical profile of \( N \) and the forcing terms, where the effect of topography vanishes also around 300 m.

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